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Ice -Core Pollen Studies From the Dunde and Guliya Ice Caps, Qinghai -Tibetan Plateau, China.

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ICE-CORE POLLEN STUDIES FROM THE DUNDE AND GULIYA ICE CAPS, QINGHAI-TIBETAN PLATEAU, CHINA

A Dissertation

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ABSTRACT

This research provides the first systematic and high-resolution ice-core pollen records from two ice caps in the northern Qinghai-Tibetan Plateau. To facilitate the climatic interpretation of the pollen data from the ice cores, a 30-year pollen record from the Dunde ice core is used to establish the quantitative relationship between pollen and modern climatic parameters.

A 2000-year pollen record from the Dunde ice core reveals decadal and century time scale climatic changes. Three dry periods (i.e. A.D. 280-640, 1200-1370, and 1926-1986) and three wet periods are identifiable at Dunde. These results are compatible with other proxy data such as historical records, tree-rings, and other ice-core data from the Qinghai-Tibetan Plateau and surrounding regions. Pollen data also suggest that the early Medieval Warm Period (9th to 12th century) was warm and humid, while the late Medieval Warm Period (13th to 14th century) was warm and dry at Dunde. The cold and wet period between A.D. 1650 and 1926 is probably correlated with the Little Ice Age. An 88-year cycle of pollen concentrations is probably correlated with the 84-year Gleissberge cycle of solar activity, while an 110-year cycle of oxygen isotope variation may be

associated with the 110-year cycle of thermohaline circulation of the ocean.

In the Guliya ice core, the pollen record reveals several vegetational and climatic changes over the last 12,000 years. Before 11,500 BP, high percentages of Chenopodiaceae pollen suggest dry conditions. Between 11,500 and 9500 BP, an increase in *Artemisia* pollen suggests a relatively more humid condition than before. From 9500 to 6500 BP, a significant increase in Gramineae pollen indicates the development of a grass-dominated steppe, probably reflecting increased precipitation due to a stronger summer monsoon. Gramineae pollen decreased after 6500 BP, indicating return to drier condition. After 1900 BP, the pollen assemblage is similar to the modern ones, suggesting that the modern arid climatic regime has been established since then. The climatic history inferred from this ice-core pollen study is consistent with the pattern reconstructed from other proxy records in the Qinghai-Tibetan Plateau.

CHAPTER I

INTRODUCTION

The Earth's ice sheets and ice caps are recognized as "the best of only a few" archives of atmospheric history from which past climatic and environmental conditions may be reconstructed (Thompson et al., 1995a). In the last three decades, ice cores have provided a variety of unique climatic information, such as changes in atmospheric circulation (Legrand et al., 1991; Alley et al., 1993; Taylor et al., 1993a; Mayewski et al., 1993a, 1994, 1996), temperature (Yao et al., 1996; Dansgaard et al., 1993; Jouzel et al., 1994; Cuffey et al., 1994, 1995), aerosol loading/volcanic eruptions (Petit et al., 1990; Taylor et al., 1992; Zielinski et al., 1996, 1997), forest fires (Legrand et al., 1992; Whitlow et al., 1994; Taylor et al., 1996), anthropogenic emissions (Mayewski et al., 1990b; Thompson et al., 1988b), solar variability (Beer et al., 1990), and radionuclide deposition (Dibb, 1989; Cecil and Vogt, 1996, 1997; Naftz et al., 1994, 1996).

Most of these ice-core records have come from the polar regions. To complement these polar histories and provide further constraints upon models of global climatic variations, ice-core histories from high-elevation icce caps in the subtropics and tropics have begun to emerge. The first such history was from the Quelccaya Ice Cap in Peru (Thompson et al., 1984a, 1984b), followed by records from the Dunde Ice Cap (Thompson et al., 1988a, 1989) and the Guliya Ice Cap (Thompson et al., 1995a; Thompson et al., 1997; Yao, 1997) in western China, and the Huascaran in Peru (Thompson et al., 1995b). More recently, records were obtained from the Sajama Ice Cap in Bolivia (Thompson, 1998; Thompson et al., 1998). The latest research focuses on ice cores from the 7000 m Xixiabangma in the Himalayas (Yao, 1998).

Previous ice-core studies focus on a number of climatic proxy parameters, which can be measured from the physical and chemical properties of ice cores. However, insufficient attention has been paid to the pollen contents of ice cores. Unlike other aerosols, pollen can be related more readily to its source vegetation and can provide an independent record of vegetational response to climatic changes (Bourgeois et al., 1985; Bourgeois, 1986, 1990a, 1990b; Liu et al., 1996; Yao et al., 1997; K-b Liu et al., 1998). Ice cores from tropical and subtropical ice caps are well situated to provide a high-resolution pollen record of past climatic and vegetational changes for several reasons. First, the high accumulation rates on ice caps can yield pollen records at an annual resolution for at least the last several centuries. Secondly, compared with polar ice cores, which typically contain very few pollen, ice cores from tropical (e.g. Quelccaya) and subtropical (e.g. Dunde, Guliya) ice caps have much higher pollen concentrations and more diverse pollen types due to closer proximity to their source vegetation. Thirdly, plant communities close to the limits of their ecological requirements are the most sensitive to abrupt climatic changes. In arid and alpine environments, a small increase in moisture supply can significantly increase the density of plant cover. Therefore, ice-core pollen records from the alpine desert areas, such as the northern Qinghai-Tibetan Plateau, can provide good registration of abrupt vegetational and climatic changes.

1.1 Purpose of the Research

The purpose of the research is to provide high-resolution pollen records from two subtropical ice caps: the Dunde Ice Cap (38°06'N, 96°24'E) and the Guliya Ice Cap (35°51'N, 81°29'E) (see Figure 3.1 on page 60 for location map). The first objective is to establish the relationship between pollen and climate on the basis of instrumental observations. It is essential for interpreting the pollen variations being detected in deep ice cores.

The second objective is to reconstruct a history of high-resolution climatic changes in the Qinghai-Tibetan Plateau for the last 2000 years. High-resolution climatic reconstruction for the last 2000 years is critical for a variety of reasons: (1) times of extremes existed within the late Holocene, such as the "Little Ice Age" (Grove, 1988) and the "Medieval Warm Period" (Hughes, 1994); (2) maximum data coverage exists; (3) multiproxy reconstruction is feasible; (4) leads, lags, and rates of change within the climate system can be studied directly; and (5) potential forcing functions may be identified and tested (Thompson et al., 1993).

The third objective is to provide a Holocene high-resolution pollen record from the westernmost Qinghai-Tibetan Plateau. A pollen and charcoal record from the Guliya ice core will supplement other proxy data from the Qinghai-Tibetan Plateau to provide a detailed history of climatic changes during the last 12,000 years.

1.2 Significance and Contributions

- This is the first systematic pollen analysis from nor-polar ice cores, especially in subtropical environment.

- This study enhances our understanding of the patterns and processes of pollen transport and deposition on an ice cap, particularly in a subtropical mountain environment.

- The pollen record supplements other evidence from the Guliya and Dunde Ice Caps to provide a high-resolution record of environmental changes in the Qinghai-Tibetan Plateau for the last 12,000 years, particularly for the last 2000 years.

- The pollen data shed light on paleowind directions and air-mass movements, and help to refine regional- and global-scale circulation models and advance our knowledge of the vegetational history of this region.

1.3 Organization

The organization of the dissertation is as follows: after the introduction of Chapter 1, in Chapter 2, a review of ice core studies is undertaken, including dating methods, climate information retrieved from ice cores, and previous pollen analysis from ice cores. Chapter 3 is about the study region. Chapter 4 is on modern pollen rain. Chapter 5 focuses on establishing the relationship between pollen and climate. Chapter 6 examines the correlation between two ice cores. In Chapter 7, it focuses on reconstructing climatic changes from a pollen study of the Dunde ice core over the last 2000 years. In chapter 8, a pollen record from the Guliya ice core provides a high-resolution climatic history in the Qinghai-Tibetan Plateau. Chapter 9 is for summaries and conclusions, in which research results are summarized and directions for future research are identified.

CHAPTER II

LITERATURE REVIEW

2.1 Introduction

In 1966 the U.S. Army Cold Regions Research and Engineering Laboratory (CRREL) succeeded in drilling a core through the Greenland Ice Sheet at Camp Century. The isotopic and geochemical data from this core provided for the first time a direct window on the history of climate of the polar regions (Dansgaard et al., 1969). In the last three decades, more than 100 ice cores have been extracted (*http://www.ngdc.noaa.gov/paleo/icecore/current.html*). Physical and chemical properties studied in ice cores include: gas contents (CO_2, CH_4 etc.), stable isotopes of occluded gases, stable isotopes in ice, particles, major and trace element chemistry of ice, cosmogenic isotopes, conductivity, and other physical properties. Moreover, the success of the drilling programs suggested that very high-resolution records extending back many thousands of years had been obtained in polar and other regions (Petit et al., 1997; Thompson et al., 1998). In this review, the dating methods of ice-core study are summarized first. Secondly, the major records of climatic and environmental changes from ice cores are reviewed. In the third part, results from ice-core pollen studies are presented. Finally, previous paleoclimatic studies from the Qinghai-Tibetan Plateau are synthesized.

2.2 Dating Methods for Ice-core Studies

In recent years, ice in ice sheets and ice caps has become an increasingly important source of paleoenvironmental information. Dating from ice provides a way to establish a chronological framework. I summarize the ice-core dating methods into five groups: (1) counting of annual layers, (2) correlation with other ice cores, (3) comparison with deep sea cores, (4) radiometric dating, and (5) ice flow calculations. For determining the ages of ice cores, the first four are direct experimental tests and the fifth rests on somewhat uncertain theories.

2.2.1 Counting of Annual Layers

The basis of this method lies with looking for parameters that vary with the seasons in a consistent manner. Many chemical and physical properties are known to vary seasonally, including chemistry (Mayewski et al., 1990a; Whitlow et al., 1992; Thompson et al., 1998; Anklin et al., 1998), isotopic composition (Taylor et al., 1992; Thompson et al., 1995b), dust data (Hammer et al., 1978; Thompson et al., 1986, 1995a), and electrical conductivity (Mayewski et al., 1990a; Meese et al., 1997). Once such markers of seasonal variations are found, they can be used to find the number of years over which the ice has accumulated. A major disadvantage of these types of dating is that they are extremely time-consuming.

Chemical Parameters

Chemical parameters vary seasonally in core cores (Whitlow et al., 1992; Fuhrer et al., 1993, 1996). For example, Anklin et al. (1998) found that in the NASA-U cores from Greenland four chemical parameters exhibit seasonal variations: H_2O_2, NH_4^+, Ca_2^+, and NO_3^- (Figure 2.1). The chemical parameters have different natural sources (Fuhrer et al., 1996) so that the concentration peak of each parameter occurred at different seasons during a year. For example, Ca_2^+ has its major source from airborne soil and its peak in Greenland occurred in spring (Whitlow et al., 1992). NH_4^+ has two main sources: biological activity and biomass burning. Its concentration peaks occurred in late spring (Fuhrer et al., 1996). Therefore, the seasonality of each parameter can be used as a tool for dating. However, if during a given year accumulation is small or absent during one or more seasons, the annual signals for a species will be missing or indistinct. To resolve this problem, the best way is to use multi-chemical parameter layer for dating (Thompson et al., 1989; Fuhrer et al., 1996). Fuhrer et al. (1996) summarized the advantages that the multi-parameter approach has in estimating annual accumulation changes.

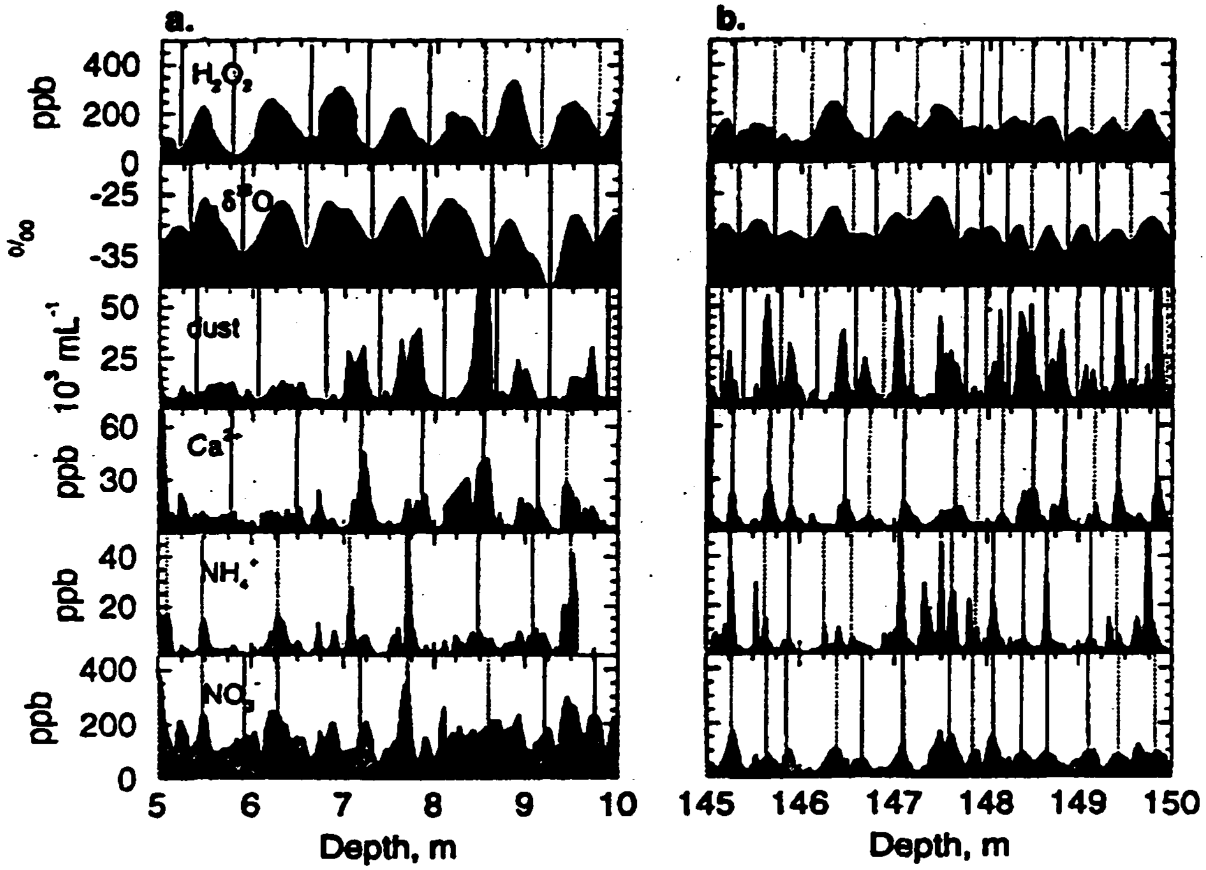


Figure 2.1 Annual layers distinguished from six markers from two different 5-m sections of NASA-U core 1. (After Anklin et al., 1998).

Isotope

Of all the isotopic parameters the most important is the ratio of ^{18}O to ^{16}O. The water molecules composed of $H_2(^{18}O)$ evaporate less rapidly and condense more readily than water molecules composed of $H_2(^{16}O)$. Thus, water evaporating from the ocean starts off $H_2(^{18}O)$ poor. As the water vapor travels towards the poles it becomes increasingly poorer in $H_2(^{18}O)$ since the heavier molecules tend to precipitate out first. This depletion is a temperature dependent process so in winter the precipitation is more enriched in $H_2(^{16}O)$ than is the case in the summer. Thus, each annual layer starts ^{18}O rich, becomes ^{18}O poor, and ends up ^{18}O rich in Antarctica (Morgan and Van Ommen, 1997) (Figure 2.2). For similar reasons the ratio of deuterium to hydrogen acts the same way. The disadvantage of using isotope is that the effects of diffusion rapidly obliterated the seasonal signal in deeper ice (Meese et al., 1997).

Dust

In some climatic regimes, atmospheric particle (dust) concentrations vary seasonally (Bull, 1971; Goudie, 1983; Ram and Illing, 1994; Isaksson and Karlen, 1994). Figure 2.3 provides an excellent illustration of annual 'dry season' dust layers in the Quelccaya Ice Cap. On the tropical Quelccaya Ice Cap, there is a consistent annual precipitation cycle during which 80 to 90% of the snow falls

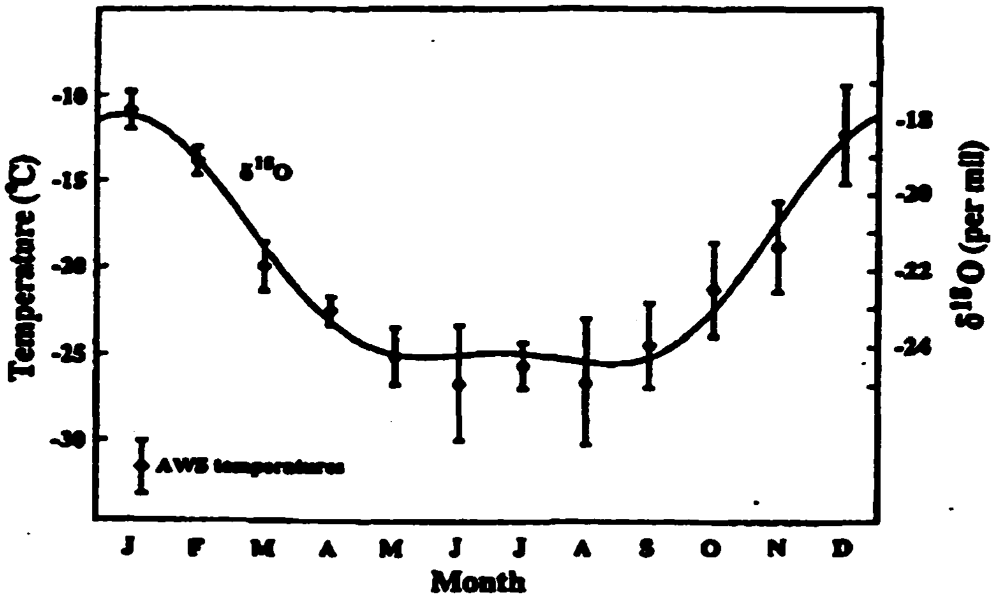


Figure 2.2 The average seasonal oxygen isotope curve of the Summit of Law Dome overlain on weather station temperature from East Antarctica. (After Morgan and Van Ommen, 1997).

Figure 2.3 Illustration of annual 'dry season' dust layers in the Quelccaya Ice Cap. (After Shimada et al., 1991).

from November to April (Thompson et al., 1984a). The distinct seasonality in precipitation results in the deposition of dry season dust bands in the Quelccaya ice cores (Thompson, 1993, 1996). Generally one dust peak is in part of a result of dust storms that occur in both Hemispheres during the spring/summer period (Ram and Illing, 1994). However, Meese et al. (1997) argued that an increased particulate concentration may not be restricted to the spring or summer and additional influxes of dust could occur during any part of the year, creating additional peaks of a non-annual nature.

Electrical Conductivity Measurements

Electrical conductivity measurements (ECM) on ice cores are strongly related to the acidity of ice (Hammer, 1980; Taylor et al., 1992; Wolff et al., 1997). The measurement is based on the determination of the current flowing between two moving electrodes with a potential difference of a few thousand volts. Strong inorganic acids cause an increase in current. Conversely, when the acids are neutralized due to alkaline dust from continental sources or from ammonia due to biomass burning, the current is reduced (Taylor et al., 1992). The lower electrical conductivity measurements occurred in the spring/summer acid peak from nitric acid production in the stratosphere (Neftel et al., 1985). Therefore, the seasonal variation of ECM has been used for ice-core dating (Yao and Thompson, 1992; Meese et al., 1997). Although ECM is an excellent seasonal indicator, non-seasonal inputs from other sources may cause additional peaks, which could be confused with the annual summer signal (Taylor et al., 1993b; Meese et al., 1997). In addition to being an annual indicator, ECM is also used for rapid identification of major climatic changes and has proved very useful in the identification of volcanic signals (Taylor et al., 1993b).

2.2.2 Correlation with Other Ice Cores

In this method one compares certain inclusions in an ice core whose age has been determined with a separate method to similar inclusions in an ice core of a still undetermined age. The major advantage of these methods is that they can be completed relatively quickly. The major disadvantage is that if the predetermined age markers are incorrect then the age assigned to the ice-core will also be incorrect.

One of the previously determined markers is typically ash from volcanic eruptions. After the eruption of volcanoes, the volcanic ash and chemicals are washed out of the atmosphere by precipitation. These eruptions leave a distinct marker within the snow. Since volcanic ash is a common atmospheric constituent after an eruption, recorded volcanic eruptions can be used to calibrate the age of ice.

This is a nice signature to use in comparing calibrated time data and an ice-core marker layer of undetermined age (Taylor et al., 1992; Meese et al., 1997). Moreover, even when absolute dates of volcanic eruptions are not available, major volcanic eruptions still can serve as reference horizons to correlate various cores (Taylor et al., 1993b).

Another approach is to compare long-term climatic changes (e.g. ice ages and interglacial warming) with markers (such as the $^{18}O/^{16}O$ ratio and atmospheric CH_4 level) found within the ice cores. For example, below 120 m in the Guliya ice core, the apparent correlation between atmospheric CH_4 levels and stadial and interstadial events inferred from $\delta^{18}O$ values in polar core cores was used to establish the Guliya time scale for the past 110,000 years (Thompson et al., 1997) (Figure 2.4). The linkage between the polar records and the Guliya record is reasonable because low-latitude moisture and temperature fluctuations likely have driven global atmospheric CH_4 concentrations, when the high northern latitudes were covered with ice and the extent of vegetation was restricted (Chappellaz et al., 1993).

2.2.3 Correlation with Deep-sea Cores

In this method, one compares certain inclusions in dated ocean cores with related inclusions found in the ice

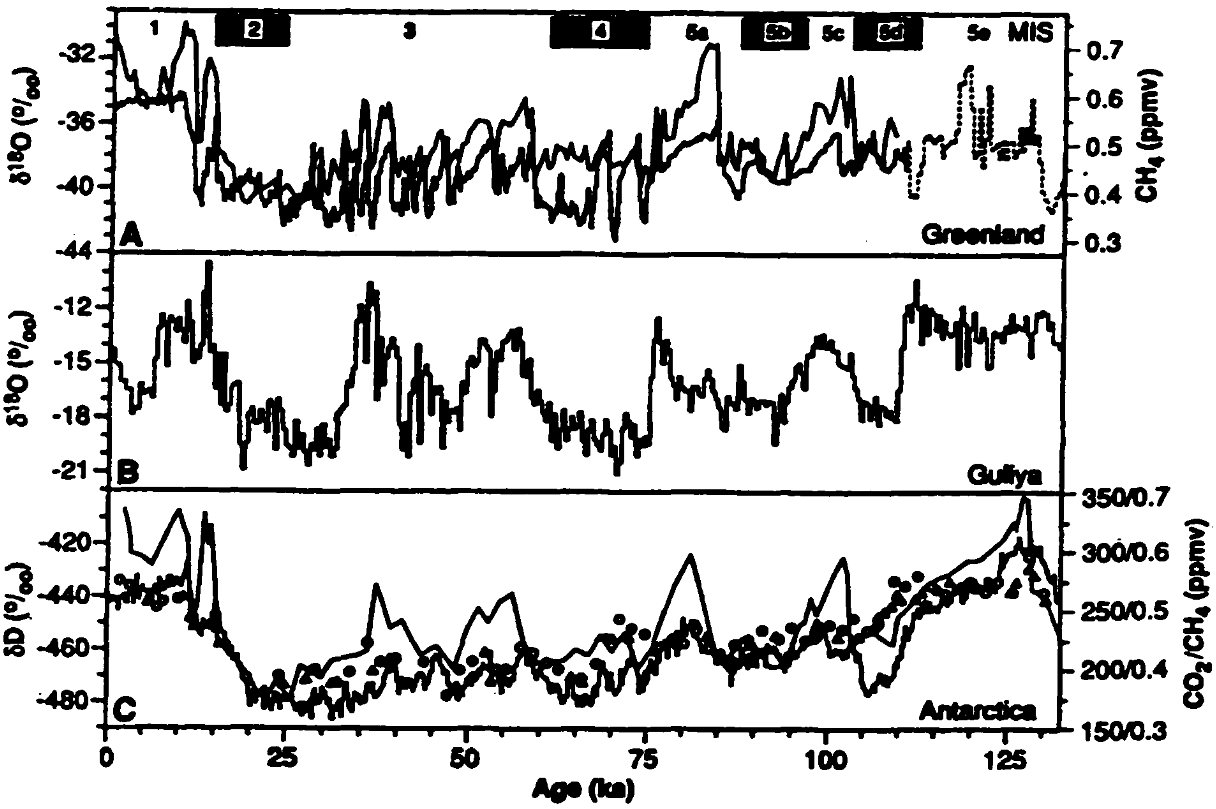


Figure 2.4 Comparison between atmospheric CH_4 levels and stadial and interstadial events inferred from $\delta^{18}O$ values in polar ice cores to establish the Guliya time scale for the past 110,000 years. (After Thompson et al., 1997).

core of a still undetermined age. Examples of such inclusions are a decrease (or increase) in temperature over a period of years that can be determined from flora and fauna found in the oceanic core and a decrease (or increase) in the ^{18}O enrichment over this same period of years (Thompson et al., 1995b; Petit et al., 1997, 1999). The major disadvantages of this method are that one must compare different signatures of climatic change that correspond to the same event and that one is not certain of the lag times (if any) between oceanic reactions and glacial reactions to the same climatic changes.

2.2.4 Radiometric Dating

Radiometric dating methods are based on the radioactive properties of certain unstable chemical elements (Lowe and Walker, 1984). Radioactive decay is time-dependent, and if the rate of decay is known, the age of the host materials can be established (Lowe and Walker, 1984). Two radiometric dating methods have been used for ice-core dating: AMS ^{14}C and ^{36}Cl.

In AMS ^{14}C dating method, it needs a quantity of glacial material from a given depth (Wilson, 1995). The materials include the gases that were trapped inside of ice or fossil that remained in ice. Wilson (1995) reported an application of AMS ^{14}C dating to polar ice core. Acceptable ^{14}C dates were obtained by dating the air entrapped in the ice cores (Andreev et al., 1986; Wilson and Donahue, 1989, 1992; Wilson, 1995; Wilson and Long, 1997). Recently, this method has been successfully used in dating the Sajama ice core (Thompson et al., 1998). The Sajama ice cores contain intact insects, insert fragments, and Polylepis bark fragments with a sufficient mass for ^{14}C dating (Thompson et al., 1998). At a depth of 130.8 m in core C-1, the corrected ^{14}C dates are 24,950±430 BP and 24,020±140 BP, respectively. Therefore, for dating the glacial in tropical and subtropical ice cores, AMS ^{14}C dating may provide a useful tool.

Another possible material which could be used for AMS ^{14}C dating is fossil pollen in ice. Especially in tropical and subtropical ice cores, they contain enough pollen grains because they are close to vegetation. However, an effective method of concentrating the pollen from an ice core has yet to be developed.

The major disadvantage of using ^{14}C dating is that a huge amount of ice must be used to gather the requisite quantity of carbon (Wilson, 1995).

Ice older than 100 ka can be dated by using ^{36}Cl, which has a half-life of 3.01 X 10^5 years (Andrews and Miller, 1980). The initial activity (^{36}Cl) is assumed to be equal to the modern (pre-nuclear testing) activity and the decay constant (λ) is 2.30 X 10^{-6} a^-. This method has been successfully used for dating the old ice of the Guliya ice core. Analysis of 27 samples showed that the ^{36}Cl activity decreases from the surface to the bottom of the Guliya ice core. The ice below a depth of 290 m is ^{36}Cl-dead, indicating that the ice is >500,000 years old, although true ages cannot be determined (Thompson et al., 1997).

2.2.5 Ice Flow Calculations

This method measures the length of the ice core, and calculates how many years have taken for a glacier of that thickness to form. First one must calculate how the thickness of the annual layer changes with depth. After this one must make some assumptions of the original thickness of the annual layer to be dated (i.e. the amount of precipitation that fell on the area in a year). This method has been used for dating several ice cores (Thompson et al., 1989, 1995b, 1998). For example, Thompson et al. (1989) used this method to date the lowest part of the Dunde ice core because the visible dust layer could only be counted to 4550 years. Using ice flow calculation, the 129.2 m depth (the LGS-Holocene transition) in the D-1 core is estimated to be about 11,950 years BP. This age is close to the annually dated age of 10,750 years ago for the LGS-Holocene transition in the Camp Century core from northern Greenland (Hammer et al., 1988). But, this is the most inaccurate of the methods used for dating ice cores because it is difficult to estimate the original thickness of the annual layer.

In summary, accurate dating is very important for ice-core studies. Without precise dating control, it is impossible to exactly reconstruct climatic changes and correlate climatic features between two sites. Usually, dating of the ice core was accomplished by using a combination of several parameters. For example, age dating of the GISP2 ice core was accomplished by identifying and counting annual layers using a number of physical and chemical parameters that included ECM, oxygen isotopic ratios of the ice, major ion chemistry, and ash from volcanic eruptions, etc.

2.3 Past Climate and the Environment Reconstructed from Ice-core Records

The detailed view of global system history revealed from ice-core properties has expanded significantly our understanding of global change (Greenland Ice-core Project (GRIP) Members, 1993; Thompson et al., 1998; Petit et al., 1999). Especially, a global change history from the ice-core studies is focused on a high-resolution view of climate (temperature, precipitation rate, air mass source, humidity, wind strength, and dustiness) and atmospheric chemistry (gases, soluble and insoluble constituents). Particular attention has been paid to the Younger Dryas, glacial/interglacial transitions, the "Little Ice Age" events, the Climatic Optimum, and rapid climatic transitions, plus detailed scans of time series to search for periodicity. With accurate dating techniques and flow modeling with the identification of global stratigraphic markers (e.g. isotopic ratios, cosmogenic events, volcanic events), these studies provide the type of perspective necessary to assess predictive global change models. In the following part, I review the climatic changes based on ice-core records from three regions: Greenland, Antarctica, and low-middle latitude regions.

2.3.1 Greenland

With the completion of the two ice-coring programs (GISP2 and GRIP) in Summit Greenland in 1993, a new era in paleoenvironmental investigation has been opened. These records are of extreme significance to our understanding of environmental change not only because they provide the highest resolution, continuous, multi-parameter view ever produced, but also because the two records can be used to validate each other (Grootes et al., 1993). In addition to providing a remarkable paleoenvironmental record, the GISP2 ice core also provides our first view of the basal conditions.

The isotopic temperature records and electrical conductivity records of GISP2 and GRIP are so similar for ice <110 kyr (Grootes et al., 1993; Taylor et al., 1993a).

However, in the bottow 10% of the cores, spanning the Eemian interglacial and previous glaciation, there are significant differences between the two records (Grootes et al., 1993). The differences between the records must result from flow deformation at one or both of the record localities (Grootes et al., 1993; Chappellaz et al., 1997).

Greenland Ice-core Project (GRIP) Members (1993) interpreted the climate proxy records from the deepest part of their core. They observed large and rapid changes in isotopic temperature with depth and concluded that these features represented rapid changes in climate during marine isotopic Stage 5e, the warmest part of the previous interglacial period (Figure 2.5).

Between 11 and 150 kyr BP, the isotopic temperature records show 23 interstadial (or Dansgaard/Oeschger) events first recognized in the GRIP record (Dansgaard et al., 1993) (Figure 2.6), and verified in the GISP2 record (Grootes et al., 1993). Intensified formation of North Atlantic Deep Water (NADW) has been invoked to explain these events (Paillard and Labeyrie, 1994; Broecker, 1997). These dramatic climate changes were not restricted to Greenland and nearby boreal areas, as evidenced by the GRIP CH_4 record (Chappellaz et al., 1993; Brook et al., 1996) (Figure 2.7). Bender et al. (1994) suggested that intensification of NADW formation resulted in some melting of ice sheets and an

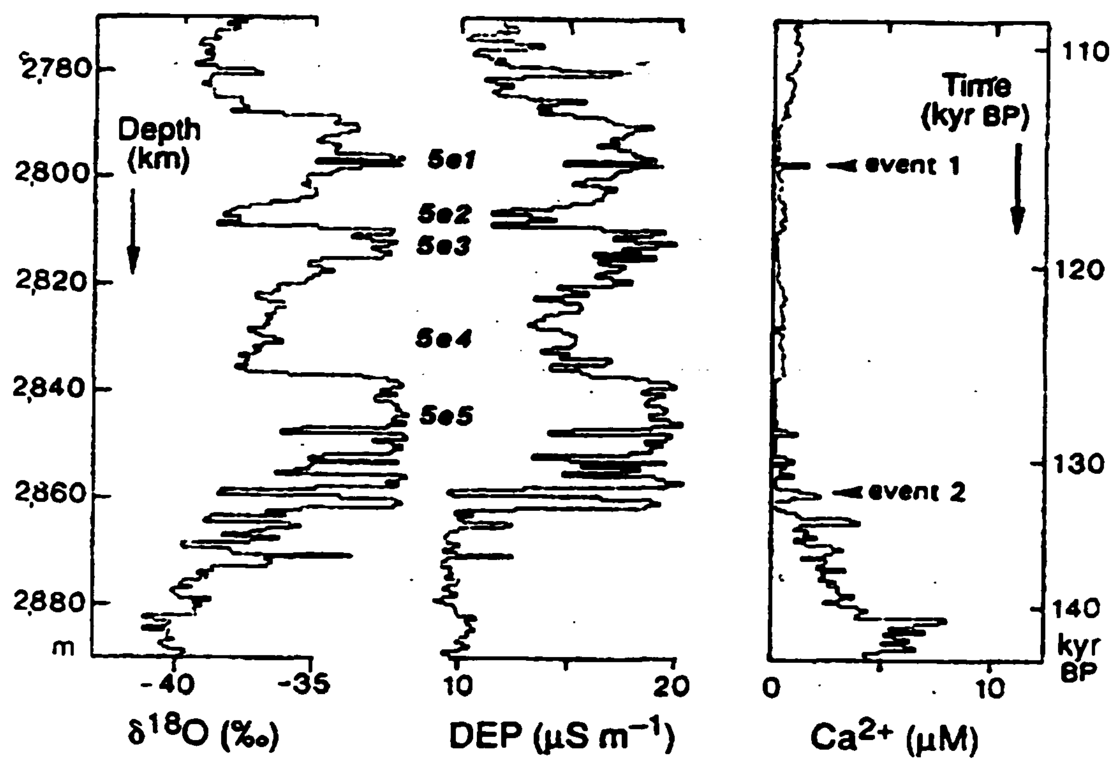


Figure 2.5 Detailed profiles of selected $\delta^{18}O$(‰), DEP and Ca^{2+} (average along 55 cm segments) through the Eemian (MIS-5e) section of the GRIP Summit core. (After Greenland Ice-core Project (GRIP) Members, 1993).

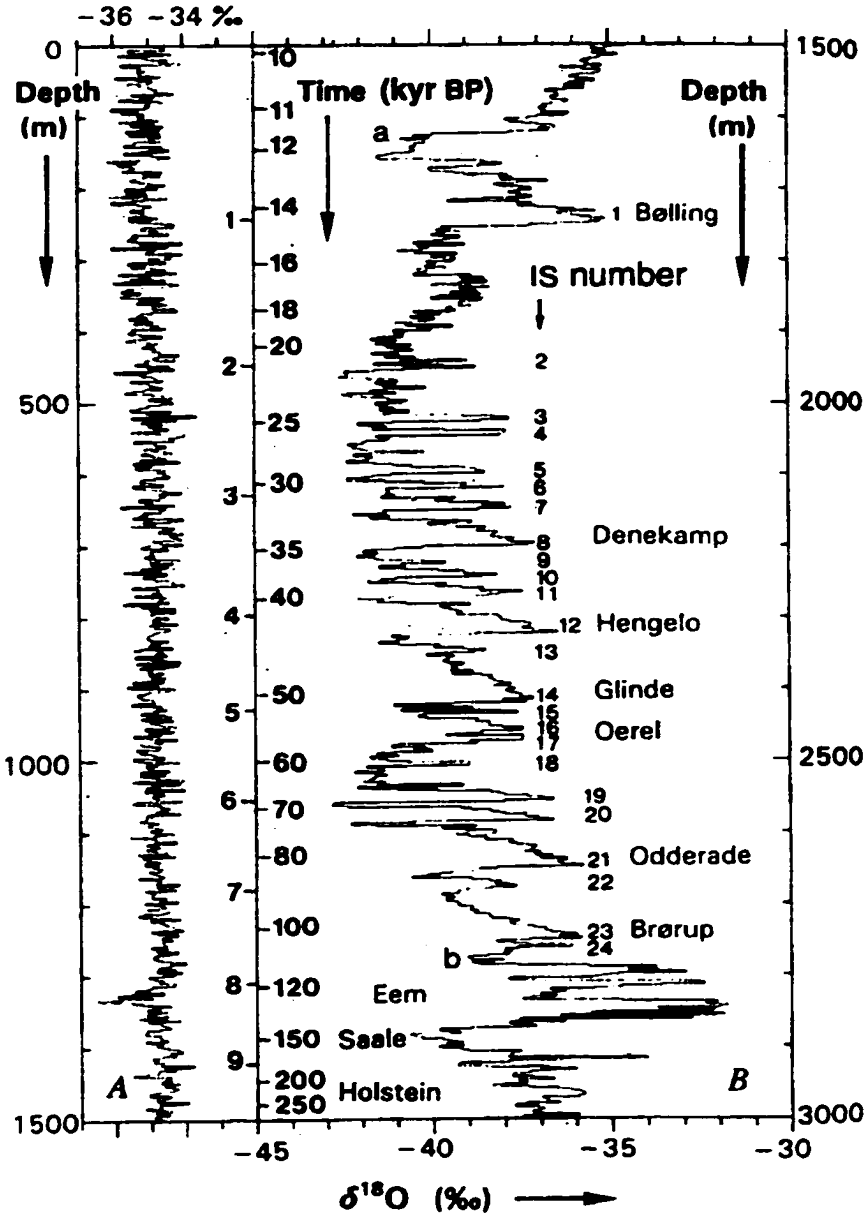


Figure 2.6 Summary of the continuous GRIP Summit $\delta^{18}O$ record. (A) from surface to 1,500 m; (B), from 1,500 to 3,000 m depth. Counting annual layers back to 14.5 kyr BP, and beyond that by ice flow modeling. (After Dansgaard et al., 1993).

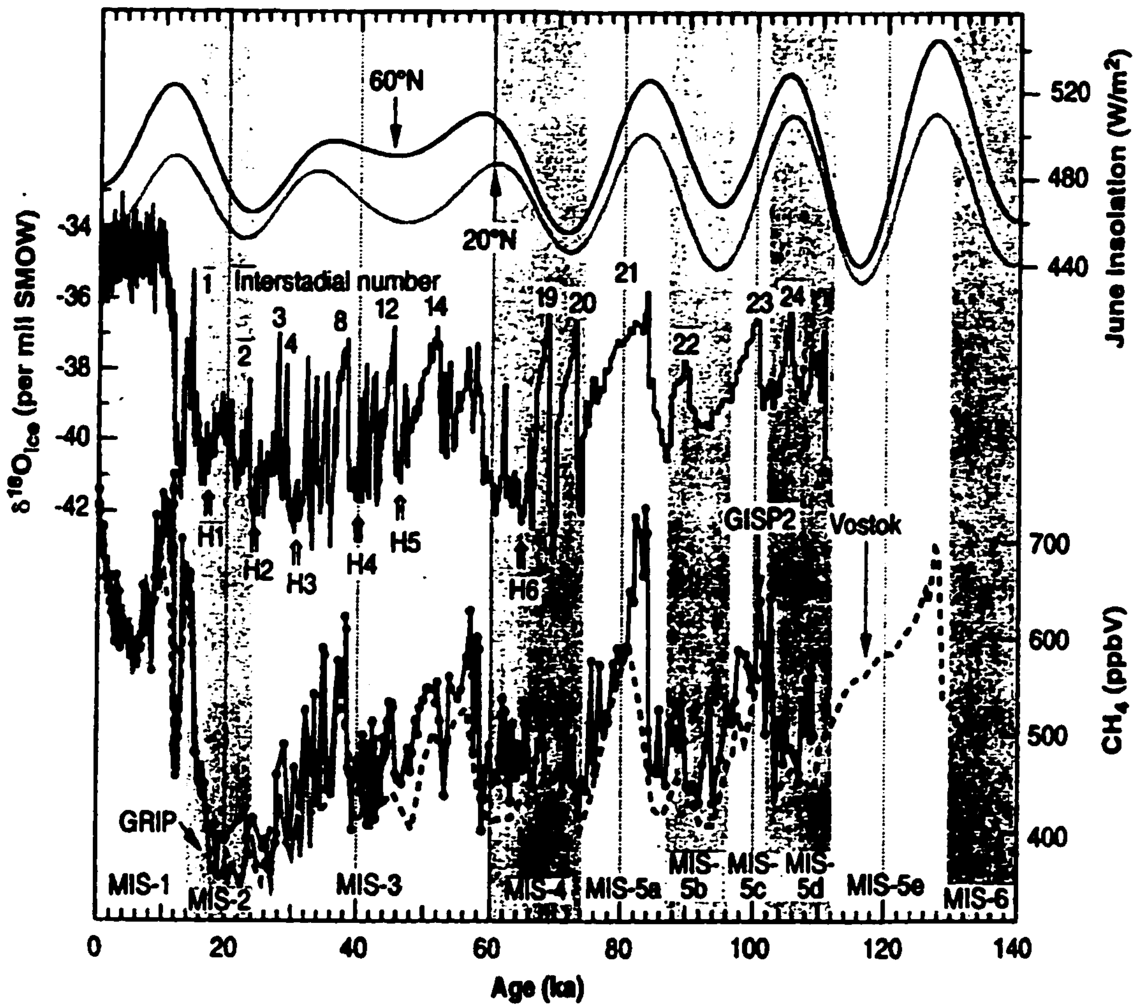


Figure 2.7 Comparison on a common time scale of GISP2 (lower solid line), GRIP (lower dotted line), and Vostok (lower heavy dashed line) methane records with Northern Hemisphere June insolation at 60°N (upper solid line) and 20°N (upper dotted line) and the GISP2 $\delta^{18}O_{ice}$ record. (After Brook et al., 1996)

increased heat flux from the Southern Ocean to the atmosphere, both of which caused Antarctica to warm. Diminution of NADW formation would eventually reverse the climate amelioration and return Greenland and Antarctica to cold, glacial conditions.

The Younger Dryas (YD) was the most significant rapid climate change event that occurred during the last deglaciation of the North Atlantic region. Ice-core studies have focused on the abrupt termination of this event (Mayweski et al., 1993) because this transition marks the end of the last major climate reorganization during the deglaciation. High resolution continuous measurements of GISP2 major anions (chloride, sulfate and nitrate) and cations (sodium, magnesium, potassium, calcium and ammonium) were used to reconstruct the paleoenvironment during the YD because these series record the history of the major soluble constituents transported in the atmosphere and deposited over central Greenland (Taylor et al., 1997) (Figure 2.8). These multivariate glaciochemical records provide a robust indication of changes in the characteristics of the sources of these soluble components or changes in their transport paths, in response to climate change. The climate change that accompanied the YD was not restricted to Greenland. The record of variations in the CH_4 concentration of trapped

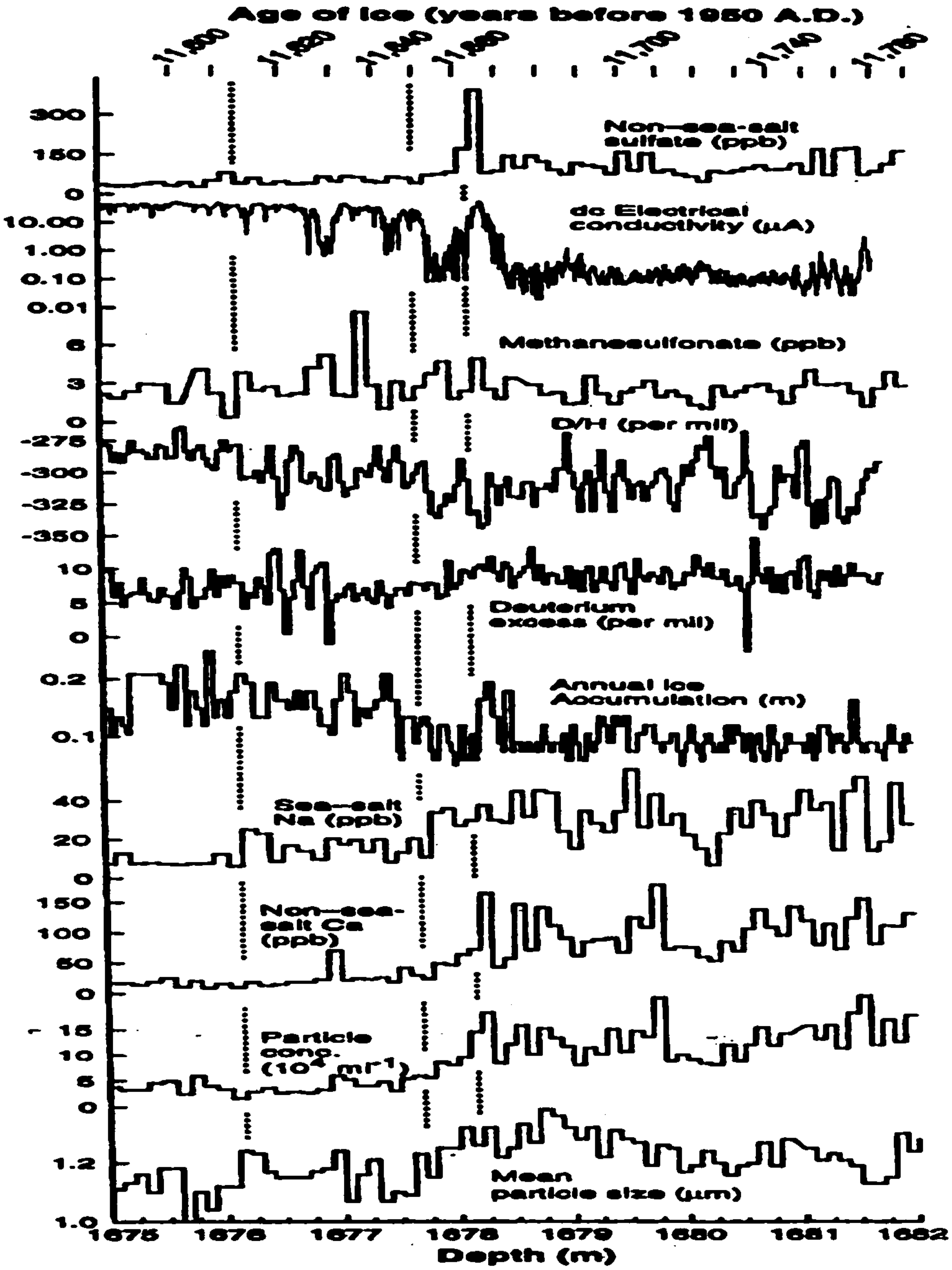


Figure 2.8 Comparison between chemical species during the Younger Dryas-Holocene transition as recorded in the GISP2 core. The dashed vertical lines are time horizons to control comparison between plots. (After Taylor et al., 1997).

gases in the GRIP ice core (Chappellaz et al., 1993) shows that tropical and subtropical climates were colder and drier during the YD. The major natural source region of CH_4 is low-latitude wetlands (Chappellaz et al., 1993). Higher atmospheric concentrations are presumably due to the greater area extent of tropical and subtropical wetlands (Chappellaz et al., 1993).

The ammonium flux record from GISP2 provides an estimate of continental biogenic source strength (Mayewski et al., 1993a) during the YD. Since ammonium concentrations are highest near continents and decrease with transport as a consequence of deposition (Mosley-Thompson and Thompson, 1994), it appears that continental sources close to Greenland (North America and Europe) were not as dramatically affected during the YD as were low-latitude wetland regions, as evidenced by the CH_4 record. This may indicate the continued importance of ice sheets and permafrost in limiting the growth of vegetation at higher latitudes until the end of the YD. Both low-latitude source CH_4 and ammonium rise at the end of the YD (Chappellaz et al., 1993; Mayewski et al., 1993b).

The Little Ice Age (LIA) and Medieval Warm Period (MWP) environments (the most recent analogs for conditions cooler and warmer, respectively, than the present century) can be characterized by interpreting the multi-parameter GISP2 series.

Accumulation rate, which is an indicator of transport distance from the open ocean plus temperature to deposition, is generally lower during the LIA than the MWP (Meese et al., 1994). Initial measurements of CO_2 in air bubbles of the GISP2 core (Wahlen et al., 1991) indicate that between A.D. 1530 and A.D. 1810 atmospheric CO_2 levels remained relatively constant at 280 ppmv. After this period, concentrations rise rather abruptly and smoothly connect to the atmospheric observations at Mauna Loa.

Dust (e.g., particles, calcium, magnesium, potassium) and species of marine origin (e.g., sodium, chloride, methanesulfonate) transported to central Greenland increased during the LIA. Nitrate sources (e.g., lightning, soil exhalation) decreased during the LIA. Ammonium peaked both at the onset and at the end of the LIA. Ammonium in Summit ice cores have been interpreted as northern high-latitude biomass-burning events (Taylor et al., 1992; Whitlow et al., 1994) based on their association with other chemical products of biomass-burning (Legrand et al., 1992).

2.3.2 Antarctica

The Russian-American and French international effort for drilling in ice achieved both a technical and scientific success by reaching a depth of 3,350 m at the Russian Vostok station (78°S, 106°E; elevation 3,488 m; mean temperature -55°C) (Petit et al., 1997). The climatic records from the Vostok ice core are the longest continuous records of Antarctica. Geochemical measurements have been conducted to a depth of 3350 meters, and the results indicate that the age of the ice at this level is approximately 420,000 years, covering four full glacial/interglacial cycles (Petit et al., 1997, 1999) (Figure 2.9). Data from the upper 2100 meters (corresponding to ~160,000 years) are included in Figure 2.10. The last interglacial period (centered at 130,000 years ago) had characteristic temperatures that were slightly higher than today. Beginning about 130,000 years ago, Antarctic temperatures dropped in a step-wise fashion until they reached full glacial values of at least -6°C, relative to today, between 25,000 and 18,000 years ago. The last glacial/interglacial transition began around 18,000 years ago and ended around 10,000 years ago with a slight cooling trend apparent throughout the last 10,000 years.

The most often-cited finding from the Vostok ice core is the close correspondence between temperature and concentrations of two radiatively active gases: CO_2 and CH_4. During glacial periods, atmospheric CO_2 and CH_4 levels were 30% and 50% lower, respectively, than interglacial levels, (Barnola et al., 1987; Chappellaz et al., 1990). Because CO_2

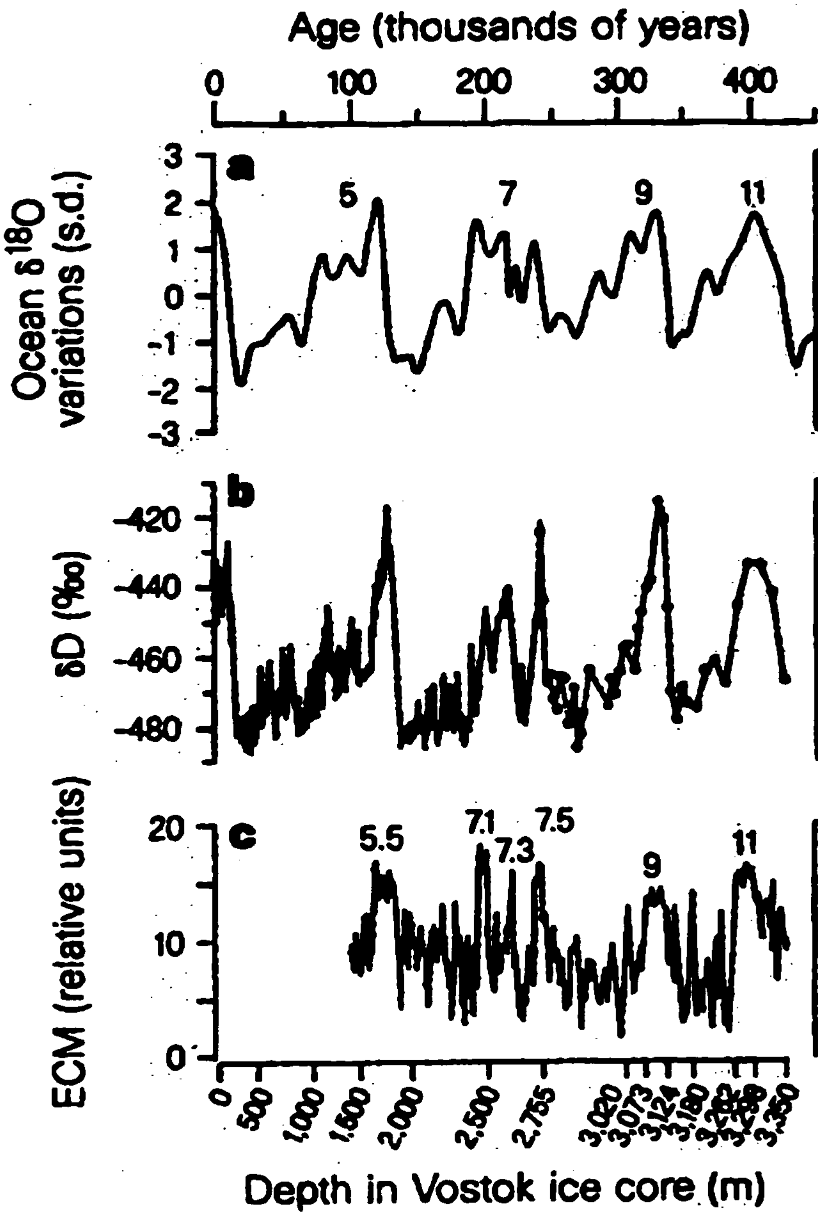


Figure 2.9 Comparison between Vostok and marine climate records over the past 400,000 years. A. Ocean ^{18}O variations from deep-sea core. B. Vostok δD values. C. Vostok ECM signal. Numbers indicate marine stages. (After Petit et al., 1997).

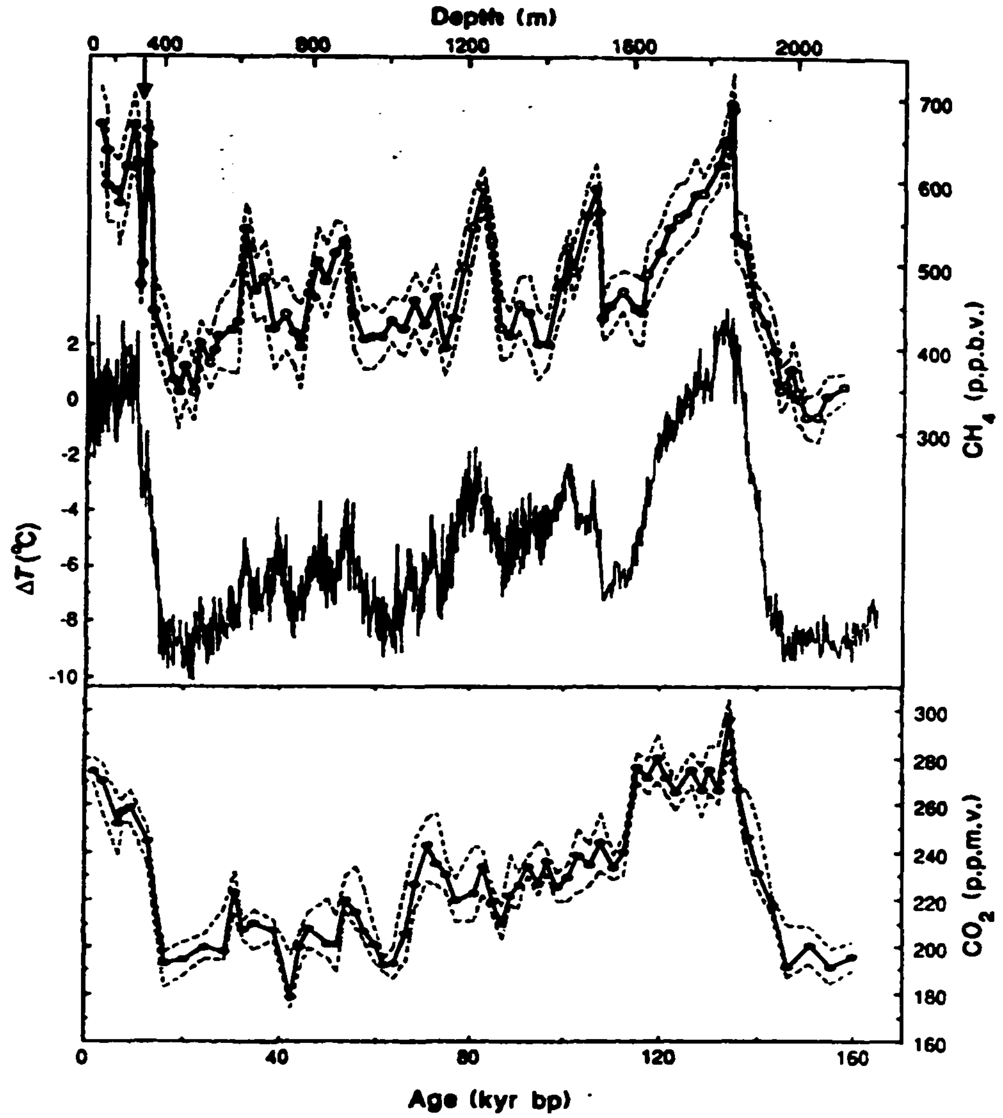


Figure 2.10 Summary of Vostok ice-core records. Upper curve: CH_4 record. Middle curve: isotope surface temperature record as a difference from the modern temperature value (-55.5°C). The lowest curve shows the Vostok CO_2 record. (After Chappellaz et al., 1990).

and CH_4 absorb a portion of the outgoing long wavelength radiation and reemit it back to the Earth's surface (a process better known as the greenhouse effect), higher concentrations of these gases in the atmosphere will tend to raise global surface temperatures (with all other factors being equal). The fact that atmospheric CO_2 and CH_4 levels were lower during the glacial period helps explain the lower temperatures and increased continental ice volume during these periods. In a study using a general circulation model to test the sensitivity of global climate to CO_2 and CH_4 variations, Lorius et al. (1990) estimated that as much as 50% of the temperature variations over the last full climate cycle could be related to observed CO_2 and CH_4 variations. The strong degree of covariation between the concentration of greenhouse gases and climate throughout the last 160,000 years has been used to predict future climate changes related to the buildup of CO_2 and CH_4 in the atmosphere as a result of anthropogenic activities.

The power spectrum of $\Delta T1$ (Figure 2.11) shows a large concentration of variance (37%) in the 100-kyr band along with a significant concentration (23%) in the obliquity band (peak at 41 kyr). This strong obliquity component is roughly in phase with the annual insolation at the Vostok site (Petit et al., 1999). This supports the notion that annual

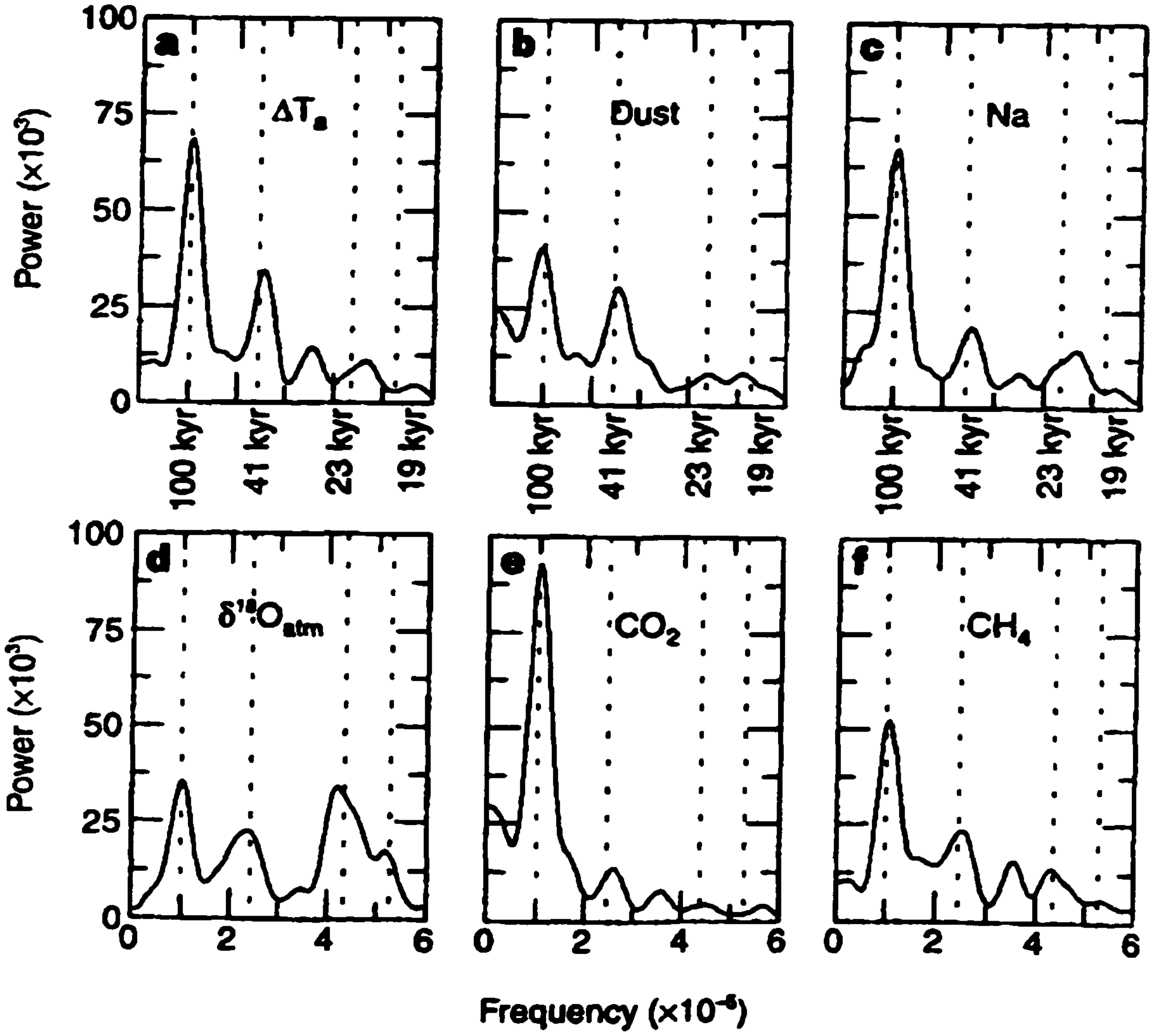


Figure 2.11 Spectral properties of the Vostok time series. Frequency distribution (in cycles yr^{-1}) of the normalized variance power spectrum (arbitrary units). Spectral analysis was done using the Blackman-Tukey method: a, isotopic temperature; b, dust; c, sodium; d, $\delta^{18}O$; e, CO_2; and f, CH_4. Vertical lines correspond to periodicities of 100, 41, 23 and 19 kyr. (After Petit et al. 1999).

insolation changes in high southern latitudes influence Vostok temperature (Lorius et al. 1990).

2.3.3 Tropical and Subtropical Ice Cores

Since the 1980s, ice-core records from tropical to subtropical areas have become a significant component of ice core studies (Thompson et al., 1984a; Thompson et al., 1995a; Thompson et al., 1997; Thompson et al., 1998). It is important to retrieve tropical and subtropical ice cores, since they fill the gap between other glacial records to the north polar region and to the south polar region. In conjunction with other proxy records, many useful data sets have been gained from tropical and subtropical ice cores, thereby providing us with a more extensive view of short-term climate variability around the globe. It is important to expand our view of climate variability by integrating the non-polar ice-core histories with those available from high latitudes. Such a suite of comparable records are essential to understanding the climate system which depends upon an historical perspective of the magnitude and synchronicity of natural climate variability and rates at which natural changes in the system have occurred. These ice-core climate records also shed light on a number of issues of great scientific concerns, such as those related to increasing CO_2, CH_4, and other trace gases, global scale events like El Niño-Southern Oscillation, the Little Ice Age, abrupt climate changes, and glacial/interglacial stage transitions. Ice cores from the tropics and subtropics have provided an invaluable record for enhancing our current understanding of the climate system and for making future predictions (Thompson et al, 1989, 1995b, 1997, 1998).

Three ice cores have been drilled to bedrock (140 m) from the Dunde Ice Cap in the Qilian mountains on the northeastern margin of the Qinghai-Tibetan Plateau. The $\delta^{18}O$, microparticle, and anion records developed from these cores provide a record of environmental change for the northern regions of the Qinghai-Tibetan Plateau for the Holocene and the late stages of the last glacial cycle (Thompson et al., 1989; 1990; Yao and Thompson, 1992). The relatively high anion and microparticle concentrations measured in ice cores from the Dunde primarily reflect the influx of dust from the large desert basins (e.g., Qaidam and Taklamakan deserts) on the northern margin of the Tibetan Plateau. The glacial to interglacial transition is marked by decreasing microparticle content, less negative oxygen isotope ratios, and higher concentration of anions (Thompson et al., 1989) (Figure 2.12). Enhanced dust deposition during the late glacial stage (LGS) may have resulted from increased wind strength during this period, consistent with data from polar ice cores. Conversely, the increasing concentration of anions in the Dunde core during

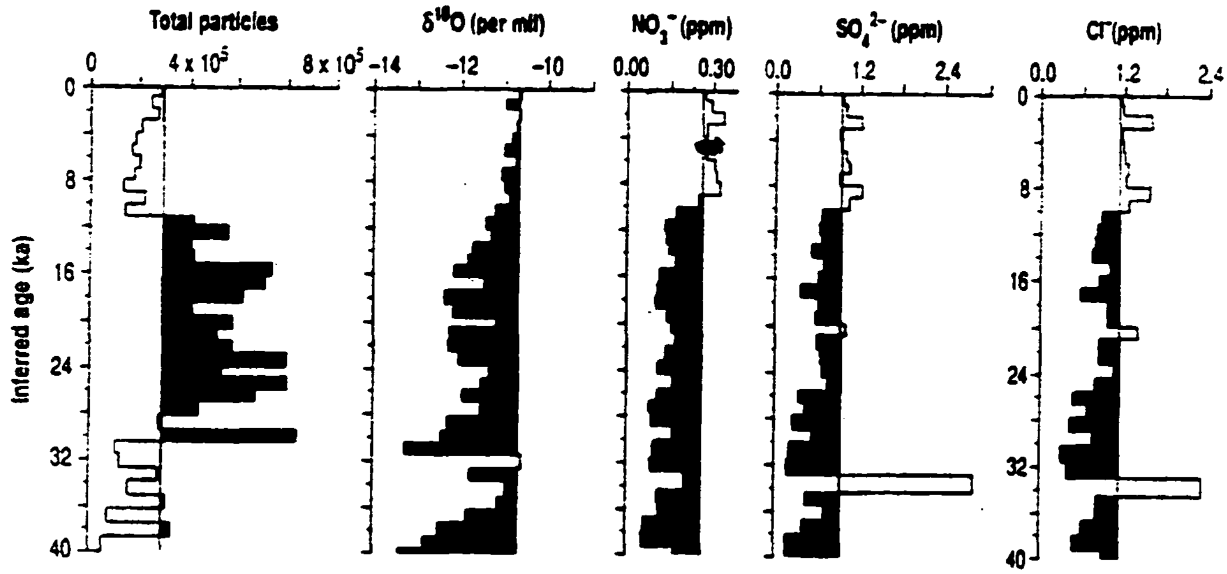


Figure 2.12 Discrete 1,000-year average of dust concentrations (diameters>=2.0 μm) and $\delta^{18}O$ in core D-1 and NO_3^-, SO_4^{2-}, and Cl^- in core D-3 for the last 40,000 years at Dunde. The average values are for all samples in the 40,000-year record and not for the individual 1,000-year averages. There are 5 m of ice core below the 40,000-year model cutoff shown her: ka, thousand years ago. (After Thompson et al., 1989).

the Holocene suggests an increase in inputs of soluble aerosol originating from the Qaidam Basin as a result of desiccation of lakes in the region (Thompson et al., 1989).

In 1992, a 308.6 m ice core was recovered from the Guliya Ice Cap, located in the west Kunlun mountains in the westernmost Qinghai-Tibetan Plateau, China. The ice-core record provides evidence of regional climatic conditions over the last glacial cycle (Thompson et al., 1997). ^{36}Cl data suggest that the deepest 20 meters of the core may be more than 500,000 years old. The $\delta^{18}O$ change across Termination I is ~5.4 per mil, similar to that in the Huascarán (Peru) and polar ice cores. Three Guliya interstadials (Stages 3, 5a, and 5c) are marked by increases in $\delta^{18}O$ values similar to that of the Holocene and Eemian (~124,000 years ago) (Figure 2.4). The similarity of this pattern to that of CH_4 records from polar ice cores indicates that global CH_4 levels and the tropical hydrological cycle are linked (Thompson et al., 1997). The climatic and environmental information for the past 1000 years derived from analysis of this core is presented in Thompson et al. (1995a). As was found in the Dunde ice cores, the Guliya ice core revealed that the Little Ice Age was not as cold as expected. Calcium was found to be a good proxy of dust (Thompson et al., 1997). In the past, cold periods are basically characterized by high calcium concentration, and warm periods by low calcium concentration. Aside from the major ion analyses, analyses have been carried out on organic acids in snow and ice in China (Wake et al., 1993). Results from the top 35 m of the Guliya ice core suggest that organic acids provide a record of the seasonal change in vegetation and source of moisture. A comparison between the Guliya ice core accumulation record and the Quelccaya ice core (Peru) accumulation record appears to indicate an excellent teleconnection on a large spatial scale (Thompson et al., 1995a) (Figure 2.13).

Ice cores recovered from the Peruvian Andes have provided evidence connecting the rise and fall of coastal and highland human activity to regional climatic changes. Annual accumulation records reconstructed from the Quelccaya ice cores (Thompson et al., 1985) indicate that an extended period of relatively high accumulation from about A.D. 750 to 1050 roughly corresponds to the time when southern highland cultures flourished (Thompson, 1992). Low accumulation intervals before and after this period in the ice core are synchronous with the flourishing of Peruvian coastal cultures, which declined during highland wet periods. Similarly, the highland cultures declined during the low accumulation intervals seen in the ice cores. An explanation may be related to ENSO events, during which precipitation today in coastal Peru and Ecuador is out of

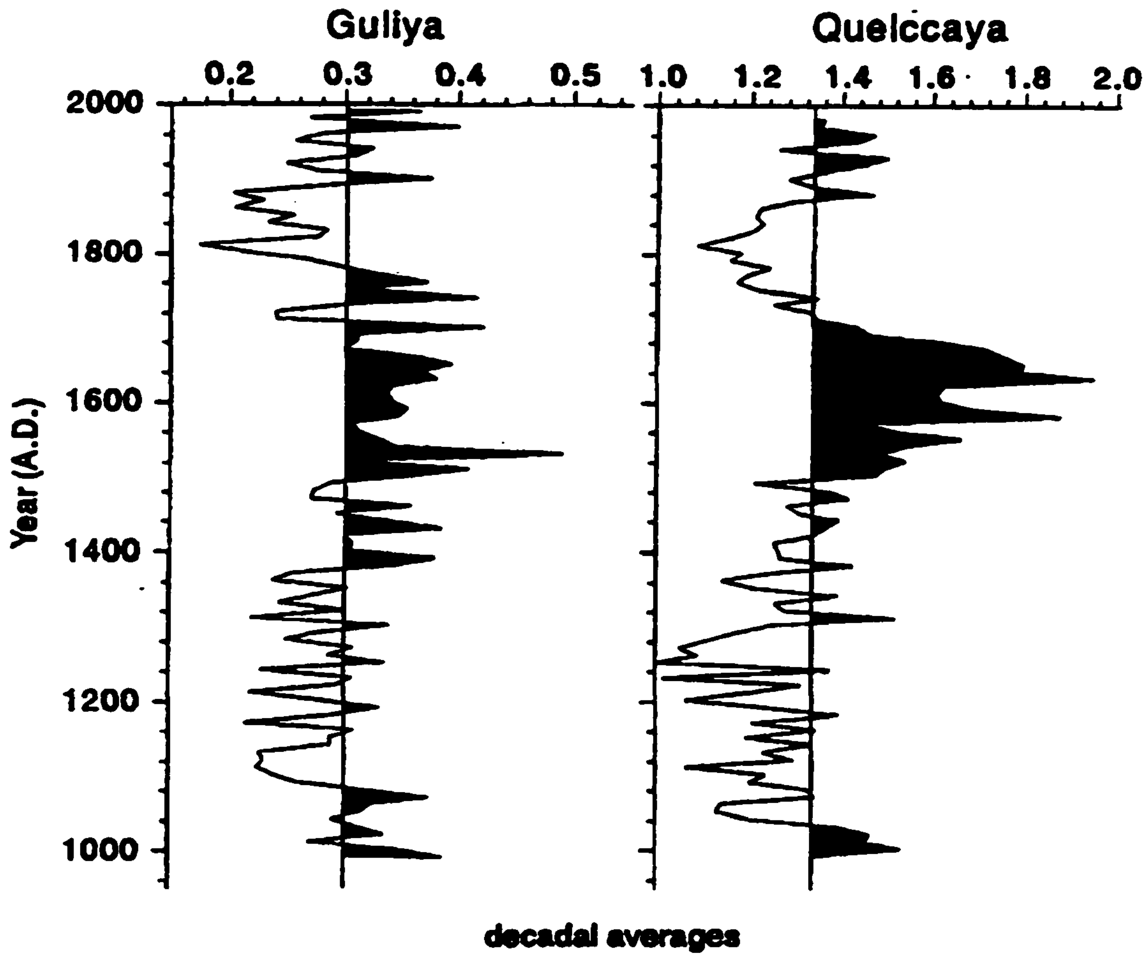


Figure 2.13 The 1000 year record of net accumulation (as decadal averages) on Guliya reveals contemporaneous trends with net accumulation on the Quelccaya ice cap, Peru. (After Thompson et al., 1995a).

phase with that in the southern Peruvian highland. The Quelccaya records suggest that this seesaw relation may have been a persistent feature of the regional climate, extending over at least a millennium. Pronounced dust peaks from the Quelccaya record lasting about 130 years and centered at about 600 and 900 A.D. seem to reflect local agricultural activity, rather than climatic or volcanic events (Thompson et al., 1988b). These ice core records document an intimate connection between climate and human activity in this region and enhance our understanding of the evolution of human society.

In 1993, two ice cores, at the depths of 160.4 m and 166.1 m, respectively, were drilled to bedrock in the Huascaran Ice Cap. The Huascaran ice-core record extends well into the LGS, a time when conditions were much colder, the atmosphere was much dustier, and biological activity in the Amazon Basin to the east was substantially reduced. Thompson et al. (1995b) suggested that the tropics were extremely sensitive to the colder LGS conditions, 5° to 8°C cooler during the LGS.

Two ice cores, at depths of 132.4 and 132.8 m, respectively, were recovered from the Sajama Ice Cap in Bolivia in 1997. One-hundred-year averages of the 25,000 year (25 kyr) Sajama record (Figure 2.14) provide a high-resolution view of the LGS (11.5 to 25 kyr). Cold and wet

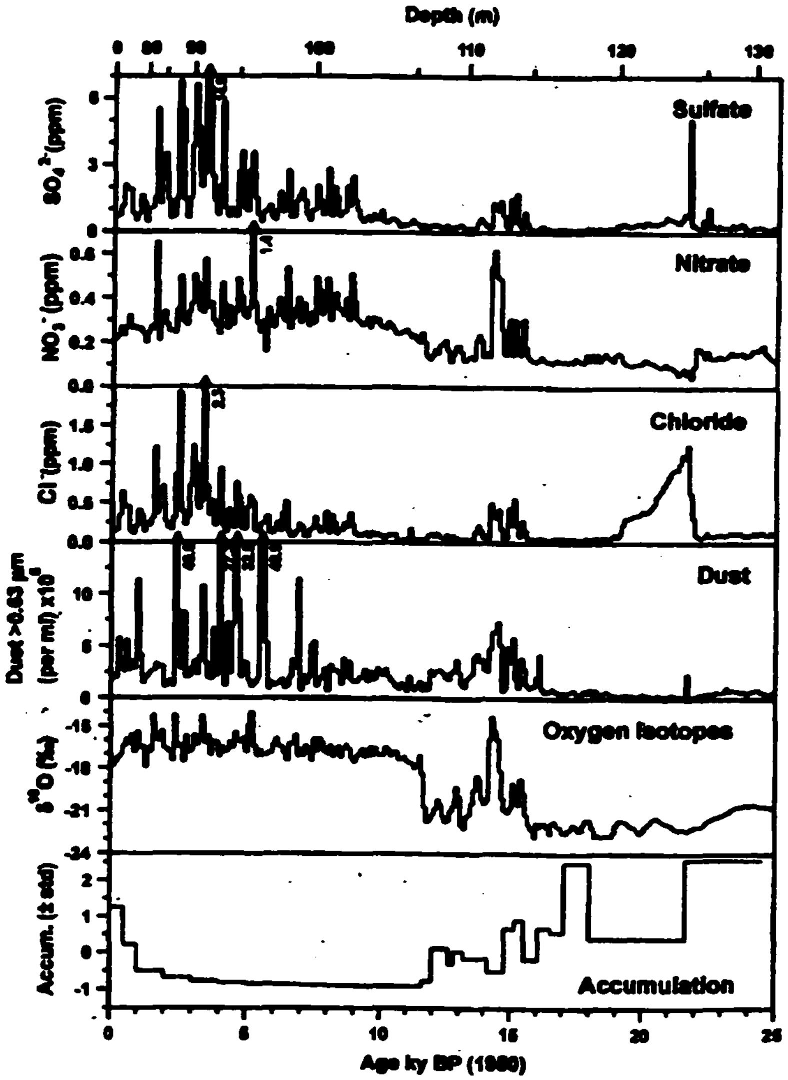


Figure 2.14 The 100-year averages of $\delta^{18}O$; insoluble dust; and Cl^-, NO_3^-, and SO_4^{2-} concentrations from C-1 in the Sajama ice core are shown for the past 25,000 years. (After Thompson et al., 1998).

conditions from 25 to 22 kyr are indicated in the Sajama record by depleted $\delta^{18}O$ value, low anion concentration, and a high regional precipitation of dust. The large increase in Cl^- and large decrease in net accumulation at 22 kyr indicate abrupt changes to a dry climate. A deglaciation climatic reversal (DCR) began at 14 kyr as the climate shifted abruptly to colder conditions that were similar to those attributed to the Younger Dryas stadial. The colder conditions of the deglaciation climatic reversal persisted until 11.5 kyr, when a sudden warming occurred within a few centuries, marking the onset of the Holocene. The comparison of the Sajama ice cores with other tropical, subtropical, and polar records shows large-scale similarities and important regional differences (Thompson et al., 1998) (Figure 2.15).

2.4 Previous Pollen Studies from Ice Cores

2.4.1 Polar Ice Cores

A number of papers have been devoted to palynological investigations of ice cores from the Canadian Arctic (Fredskild and Wagner, 1974; McAndrews, 1984; Bourgeois, 1986). These studies have demonstrated the possibility of correlation between pollen fluctuations and glaciological events in the high Arctic. Long-distance transported pollen grains generally dominate the pollen spectra, although the

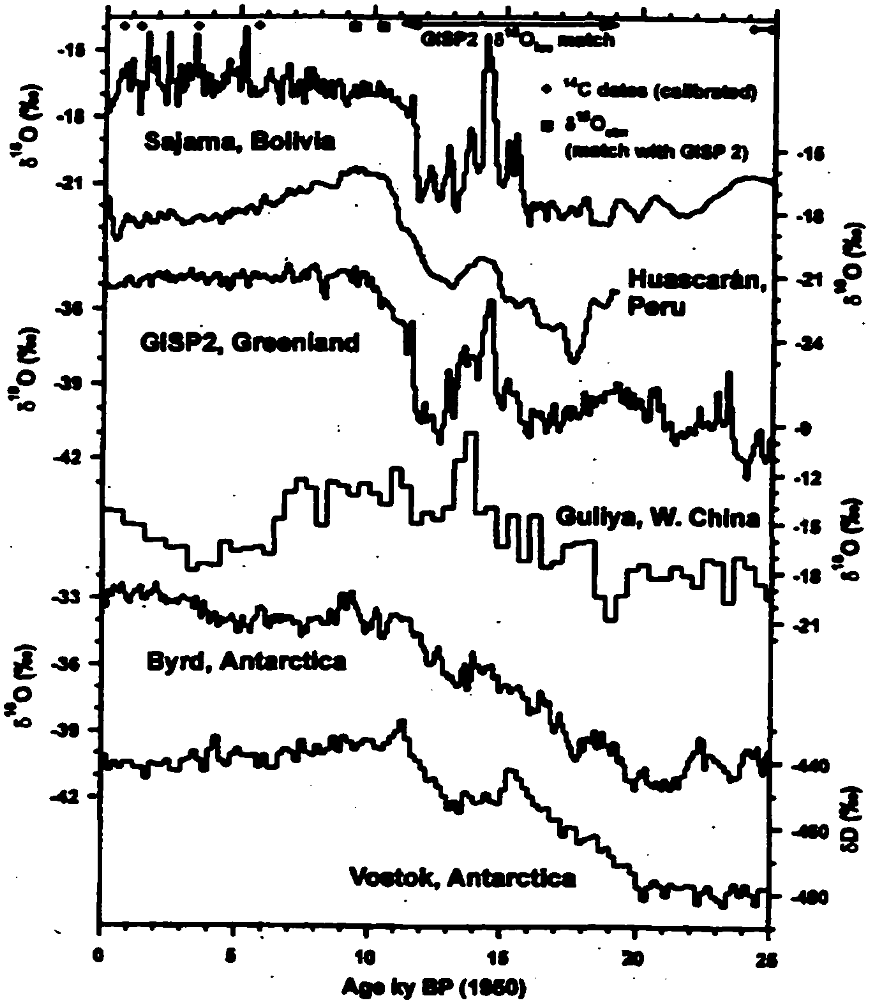


Figure 2.15 The global extent of the LGS and a climatic reversal (cooling) during deglaciations as illustrated by the stable isotope records from two tropical sites, two Northern Hemisphere sites, and two southern Hemisphere sites. All records shown are 100-year averages, except records for Vostok (200-year averages) and Guliya (400-year averages). (After Thompson et al., 1998).

presence of a considerable number of pollen grains from the local vegetation is noticed by all the authors.

Pollen changes in these Arctic ice cores have been interpreted in terms of climatic changes and changes in the local and regional landscapes (McAndrews, 1984; Bourgeois, 1986). Understanding the process of pollen deposition on ice caps in relation to modern climate is essential for interpreting the pollen variation detected in deep ice cores. The pollen content of glacial ice was first investigated by Vareschi (1935) and subsequently by Ambach et al. (1966) in firn samples from European Alpine glaciers and by Bright (1981) in snow sample from western United States. Bourgeois et al. (1985) studied pollen deposition in surface snow and shallow ice cores at fourteen sites in the Canadian Arctic Islands and on the Arctic Ocean. Their study, which concentrated on exotic tree pollen, showed a steep exponential decrease in deposition rates between the North American tree line and the High Arctic, and a flat distribution within the High Arctic and on the Arctic Ocean to the north. However, no detailed sampling was done determine when the pollen grains were being deposited on the ice cap and under what type of atmospheric conditions.

Seasonal pollen variations in snow and firn of Arctic Ice Caps were investigated by Lichti-Federovich (1974, 1975b) on the Devon Ice Cap and by Short and Holdsworth (1985) on the Penny Ice Cap. The Devon Ice Cap study covered one year of accumulation. It showed a relatively high proportion of exotic tree pollen in the winter/spring layers while the summer/autumn layers were dominated by pollen of probable local or regional origin. Contrary to the Devon Ice Cap study, the samples from the Penny Ice Cap (Short and Holdsworth, 1985) were dominated by exotic tree pollen, with the largest influx often occurring in the summer samples while the winter samples were characterized by more regional pollen types. Seasonal and annual variation of pollen content in the snow was studied by Bourgeois (1990b) from a Canadian high Arctic Ice Cap. The regional pollen types show a large seasonal variation as well as large annual variations.

McAndrews'(1984) work on the 299-m long ice core from the Devon Island Ice Cap yields a continuous pollen profile spanning the last 130,000 years. The sampling intervals ranged from 100 to 1000 years for the late Holocene (5000-100 BP), but increased exponentially to 5000 years, 50,000 years, and 70,000 years per sample in the lowermost three samples that represented the early Holocene (5000-10,000 BP), Wisconsin (10,000-60,000 BP), and the last interglacial (60,000-130,000 BP), respectively. Most of the pollen and spores were exotic. The late Holocene and last interglacial pollen assemblages were dominated by *Alnus*, whereas the early Holocene and

Wisconsin assemblages were dominated by *Betula* and *Artemisia* (Figure 2.16).

Bourgeois (1986) studied a 338 m ice core from the Agassiz Ice Cap. Low exotic pollen concentration occurred in the early Holocene and a substantial increase in pollen concentration occurred at 7600 BP. The gradual increase of pollen of regional origin at about 6000 BP could indicate that there was an expansion of vegetation. The exotic pollen concentration also increased at about 3000 BP (Figure 2.17). This increase of exotic pollen grains was unexplained (Bourgeois, 1986).

Recently, a 400 kyr old pollen record was produced from the Vavilon ice core in the Severnaya Zemlya Archipelago, Russia (Andreev et al., 1997). The majority of pollen identified in the Vavilov ice core belongs to the local flora, especially in the interval from 4.71 to 40.05 m depth. Exotic pollen, which suggests very long-distance transport, is of great interest in terms of air mass changes. For example, the presence of considerable amounts of the West-European lime species, *Tilia cordifolia* (platyphyllos), in the upper 65 m depth suggests that summer air mass dominance has been from the south-west during the last 500 years. The pollen spectra from the Vavilov ice core are completely different from the published pollen diagrams from adjacent areas of Siberia (Andreev et al., 1997).

Figure 2.16 Pollen percentage diagram from the Devon ice core. The black intervals have a sum ranging from 121 to 1271. The stippled intervals have a sum ranging from 23 to 64. (After McAndrews, 1984).

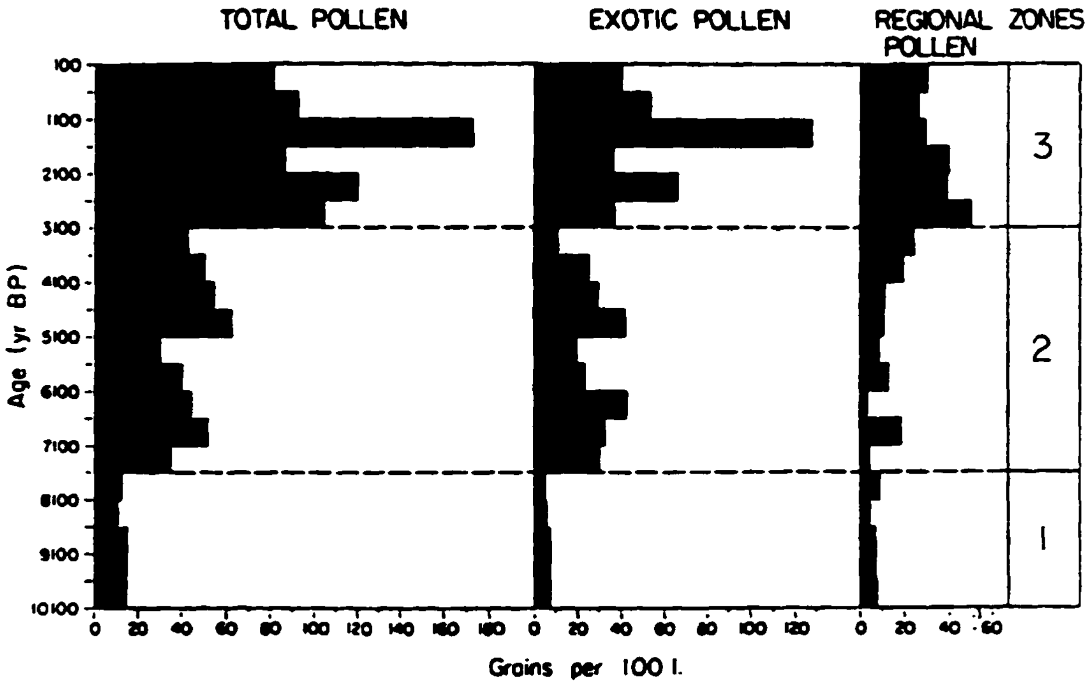


Figure 2.17 Summary of pollen concentration diagram from the Agassiz ice cap presented on a linear time scale. Each level represents the average pollen concentration per 100 liter during a 500 year period. (After Bourgeois, 1986).

2.4.2 Tropical and Subtropical Ice Cores

Pollen studies in middle to low latitudes have been reported (Liu, in Thompson et al., 1988b; Liu, in Thompson et al., 1995b; K-b Liu et al. 1998). Pollen record from the Huascaran ice core developed a record of vegetation and climate change for the Peruvian highland. Two late Holocene samples are dominated by *Alnus* (46 to 70 percent), accompanied by minor taxa such as Gramineae, Chenopodiaceae, Compositae, and several others. The high abundance of *Alnus* implies that the pollen was transported long distance and upslope from the east, since, today, *Alnus* is a small tree growing in the subpuna shrubland and dry montane forest zones between 2800 and 3900 m on the eastern Andean slopes. In contrast to the Late Holocene samples, abundances of *Alnus* in early Holocene samples are low. The samples from LGS ice contains little pollen and thus suggests that the environment was cold and the surrounding area contained sparsely vegetated superpuna or periglacial desert where pollen productivity was low (Thompson et al., 1995b).

The pollen record from the Dunde ice core is the first systematic study of pollen in non-polar ice cores (K-b Liu et al., 1998; Liu et al., 1996). Most of the pollen deposited in the Dunde ice cores are derived from xerophytic vegetation growing in the steppe and desert regions adjacent to the Dunde Ice Cap (Figure 2.18). The pollen data from the

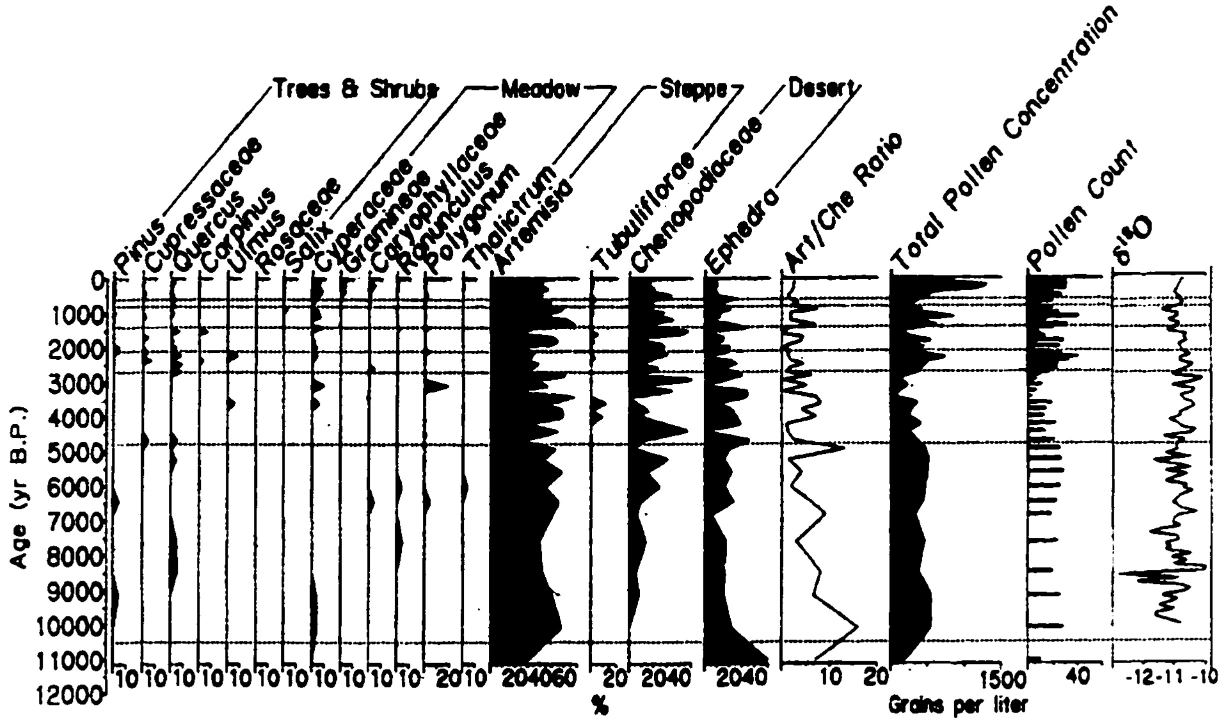


Figure 2.18 Pollen percentage diagram from the Dunde ice cores for the past 11000 years. Time resolution for samples >2000 year old is >100 year per sample. For past 2000 years, the 355 samples are grouped into 100 year averages, and pollen count is average count per sample. (After K-b Liu et al., 1998).

Dunde ice core demonstrate that the pollen record is a useful supplement to the oxygen isotope data and a sensitive paleoclimatic proxy in ice core research, because it integrates the biotic response to climatic changes at the regional scale (K-b Liu et al., 1998).

2.5 Climatic Proxy Data from the Qinghai-Tibetan Plateau

Fang (1991) suggested that, during the postglacial, lakes experienced their highest levels between 15000 and 12000 BP in central and western Tibet, and during the middle Holocene in southern Tibet and in the Qinghai basin. The pollen record from Qinghai Lake (Du et al., 1989; Kong et al., 1990) suggests a steppe or semi-desert dominated by *Artemisia* and *Ephedra* during 11000-8000 BP. The climate was warmer and more humid during 8000-3500 BP, when tree pollen increased (Figure 2.19). At Bangong Co, the stable isotope record suggests that heavy monsoon rains reached western Tibet at least from 9600 to 8700 BP, and from 7200 to 6300 BP. After 6300 BP, the overall positive $\delta^{18}O$ gradient indicates a general trend toward aridity (Figure 2.20). This is in agreement with the climatic history deduced from the detailed pollen analysis from the Sumxi Co (Figure 2.21). The two pollen records (Bangong Co and Sumxi Co) clearly show that wet conditions were established in two steps, at 10000 and 7500-7000 BP, as well as a step-wise return toward aridity from 6300 to 3800 BP. Therefore, Gasse et al. (1991) argued that

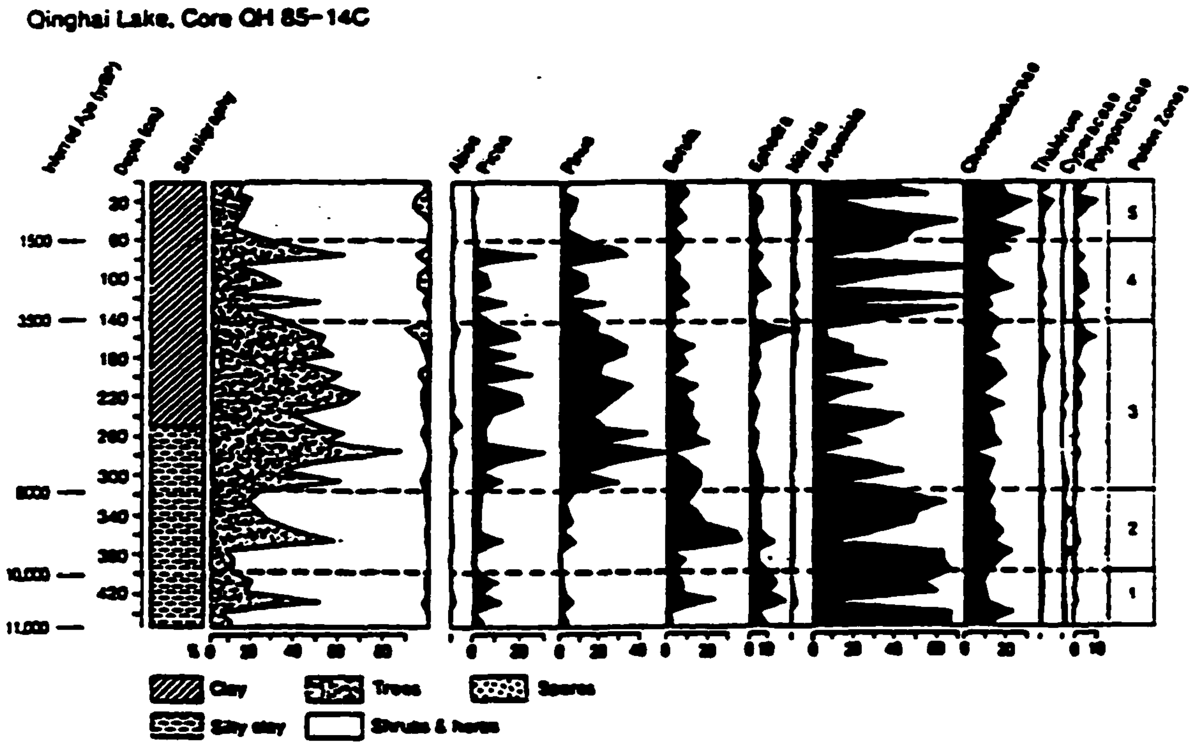


Figure 2.19 Pollen diagram from Qinghai Lake, Core QH85-14C. (After Du et al., 1989; Liu and Qiu, 1994).

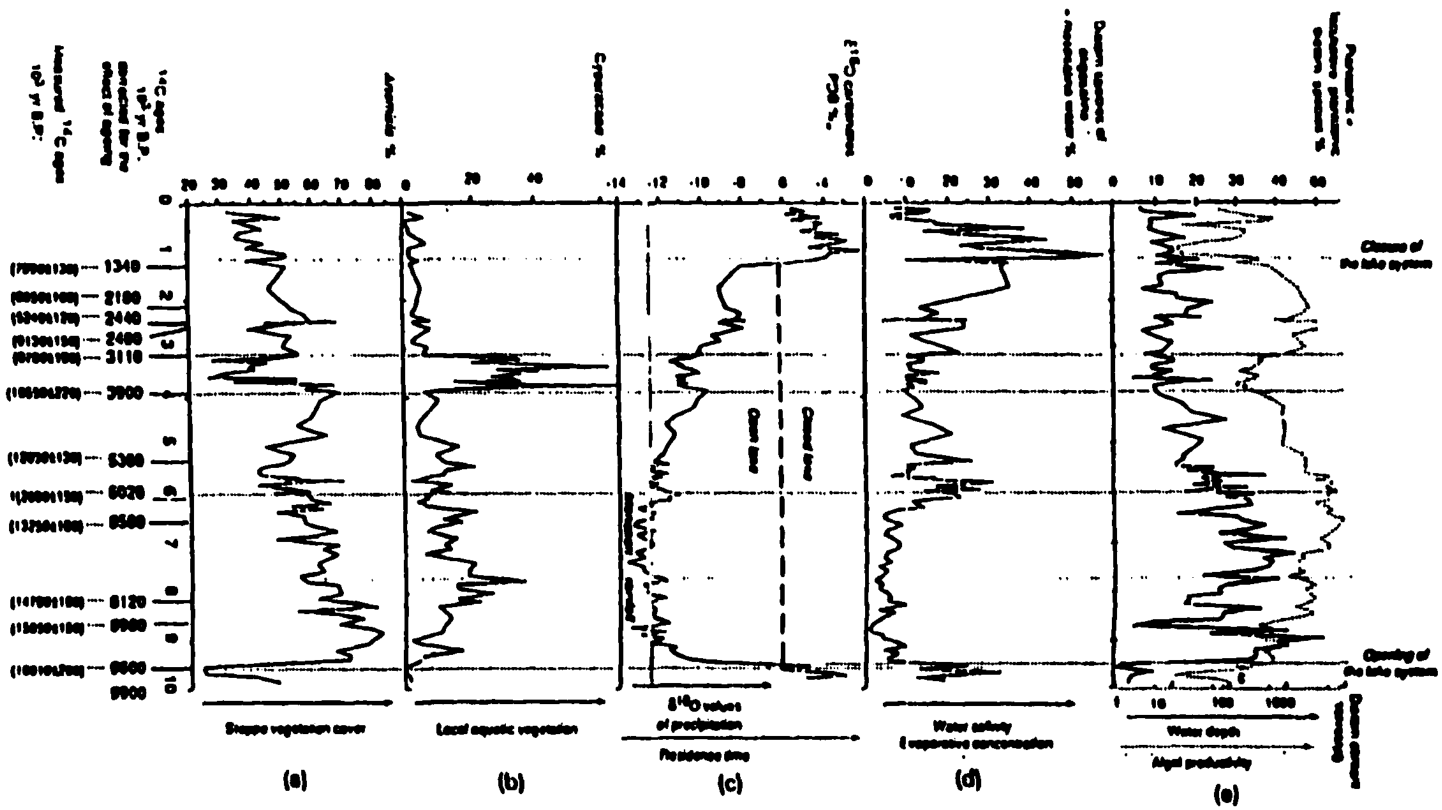


Figure 2.20 Multiproxy diagram from Lake Bangong, western Tibet. Summary results for a 12.4 m core. Measured (left) and corrected (right) AMS ^{14}C ages on carbonates (inorganic carbonates, and mollusc and ostracod shells). (a) *Artemisia* pollen frequencies, (b) Cyperaceae pollen frequencies, (c) Content of oxygen isotope in authigenic inorganic carbonates, (d) Total percentages of diatom species of oligosaline and mesosaline waters, (e) Total percentage of planktonic+facultative planktonic diatom species, and diatom content. (After Gasse and Van Campo, 1994)

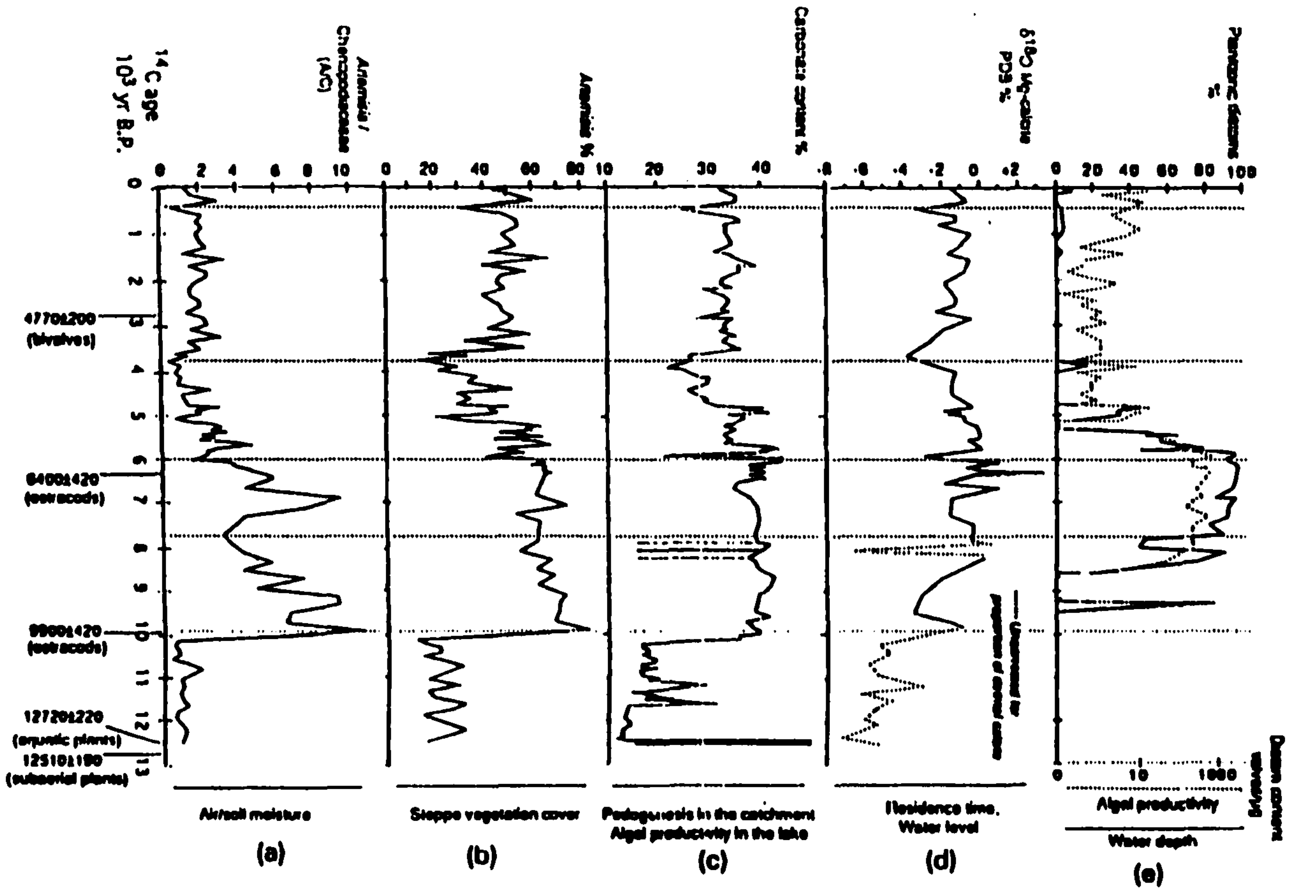


Figure 2.21 Multiproxy diagram from Lake Sumxi, western Tibet. Summary results for a 10.4 m core. AMS ^{14}C chronology, (a) *Artemisia*/Chenopodiaceae ratio, (b) *Artemisia* pollen frequencies, (c) Carbonate content, (d) Content of ^{18}O in carbonates, (e) Percentage of planktonic diatoms and diatom content of the sediment. (After Gasse and Van Campo, 1994).

these step-wise climatic changes were controlled by non-orbital and non-linear feedback mechanisms in the climate system.

Since 1987, ice-core paleoclimatic records have been produced from the Guliya and Dunde Ice Caps. Physical and chemical parameters measured from these two ice cores reveal striking climatic and environmental changes in the northern Tibetan Plateau (Thompson et al, 1989, 1997; Yao et al., 1992). The $\delta^{18}O$ records from the Dunde ice cores suggest a relatively warm interval during the mid-Holocene, about 8000 to 6000 years ago, but becoming cooler after 3000 BP. However, comparison of the two $\delta^{18}O$ histories from the Dunde and Guliya ice cores for the last 1000 years suggests that the temperature trends on these two ice caps are dissimilar at times and often in anti-phase (Figure 2.22). It is not surprising that their $\delta^{18}O$ histories appear different, particularly since moisture sources vary over the plateau. Observations of contemporary climate regimes on the plateau reveal such spatial differences (Luo and Yanai, 1983, 1984; Tang et al., 1979).

In summary, these isotopic and pollen records from the Qinghai-Tibetan Plateau will provide a baseline for correlation with the ice-core pollen records from the Dunde

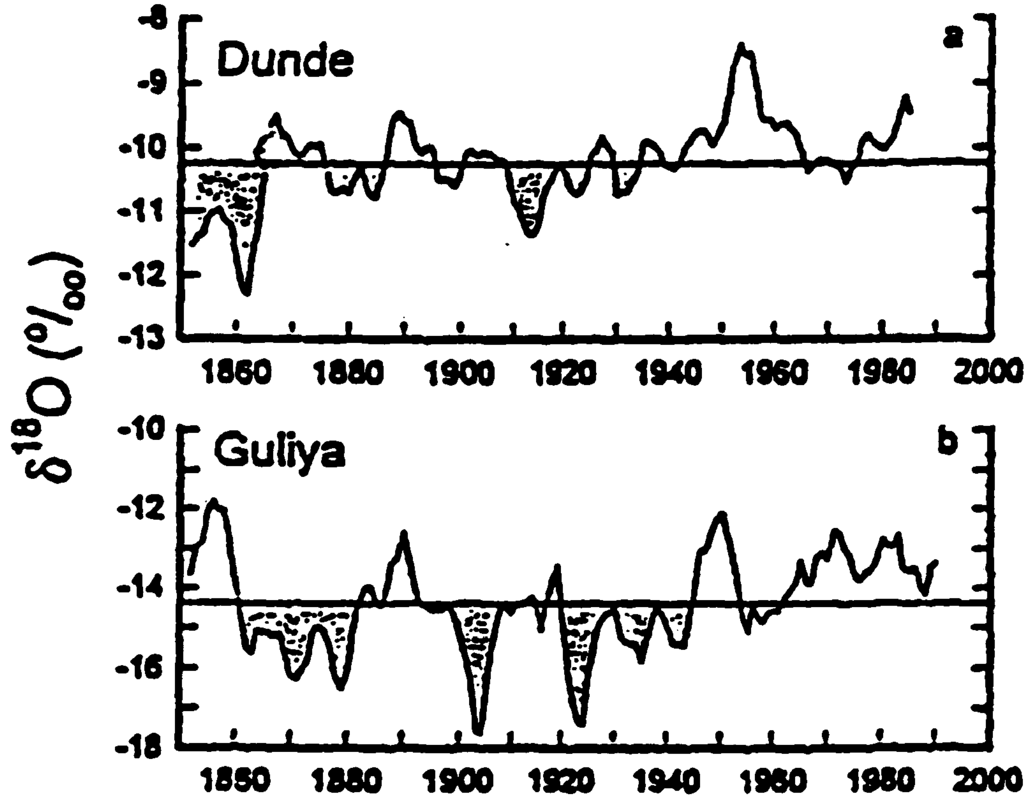


Figure 2.22 The five-year running mean of $\delta^{18}O$ from 1850 to 1992 from (a) the Dunde ice core and (b) the Guliya ice core. (After Lin et al., 1995)

and Guliya ice cores to reconstruct the regional picture of climatic and vegetational changes, especially the summer monsoon history.

CHAPTER III

STUDY REGION AND METHODS

3.1 Location

The Qinghai-Tibetan Plateau, located in western China, has a mean elevation of about 4.5 km and comprises an area half that of the United States (Zheng et al., 1979; Thompson, 1996). The Qinghai-Tibetan Plateau plays an important role in the atmospheric circulation of the Northern Hemisphere, and especially the Indian monsoon (Flohn, 1981). The sensible heat flux and the latent heat release over the Qinghai-Tibetan Plateau drive the regionally intense monsoon circulation and strongly influence global circulation patterns (Ren, 1981). Study sites from two ice caps, the Dunde Ice Cap (38°06′N, 96°24′E; summit 5325 m a.s.l.) and Guliya Ice Cap (35°17′N, 81°29′E; summit 6710 m a.s.l.), are located in the northern part of the Qinghai-Tibetan Plateau (Figure 3.1).

3.2 Climate

On the winter circulation map of China (Figure 3.2), a strong cold anticyclone over Mongolia (the Mongolian High) and a pronounced low pressure over the Aleutian Islands (the Aleutian Low) determine the winter airflow and climate over contiguous China. Prevailing winds over most parts of China

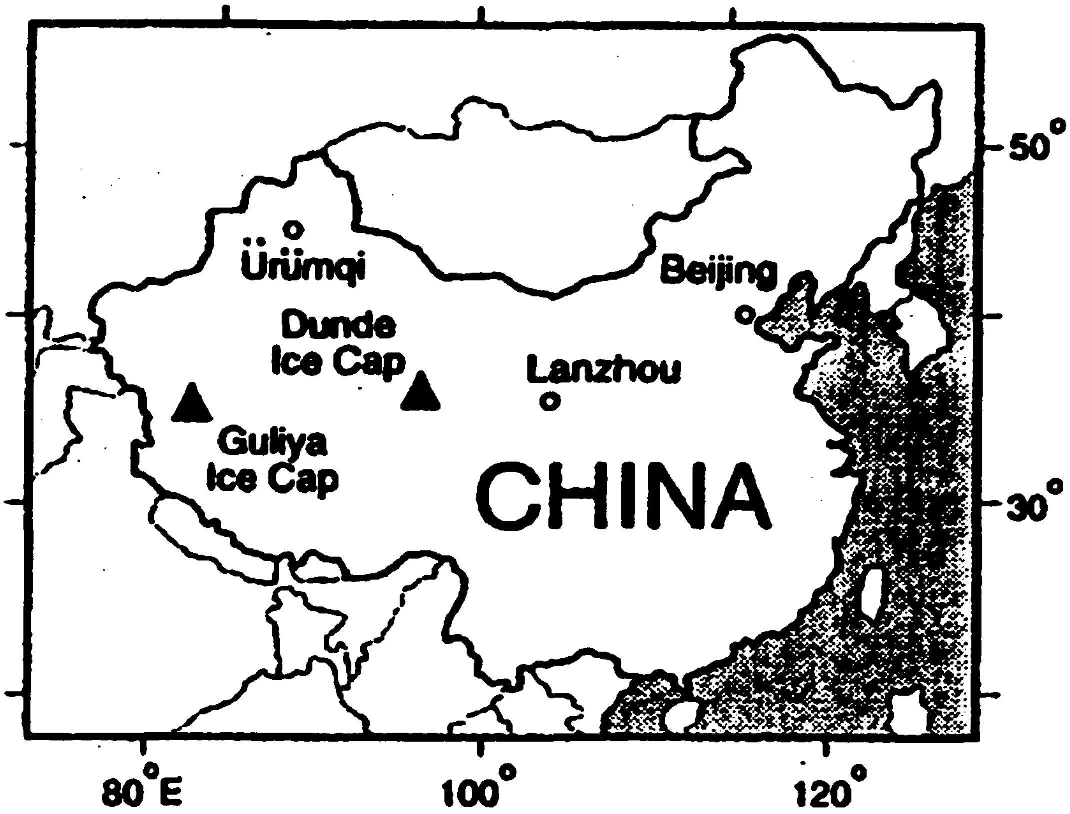


Figure 3.1 Locations of the Guliya and the Dunde ice caps. (Modified from Thompson et al., 1997).

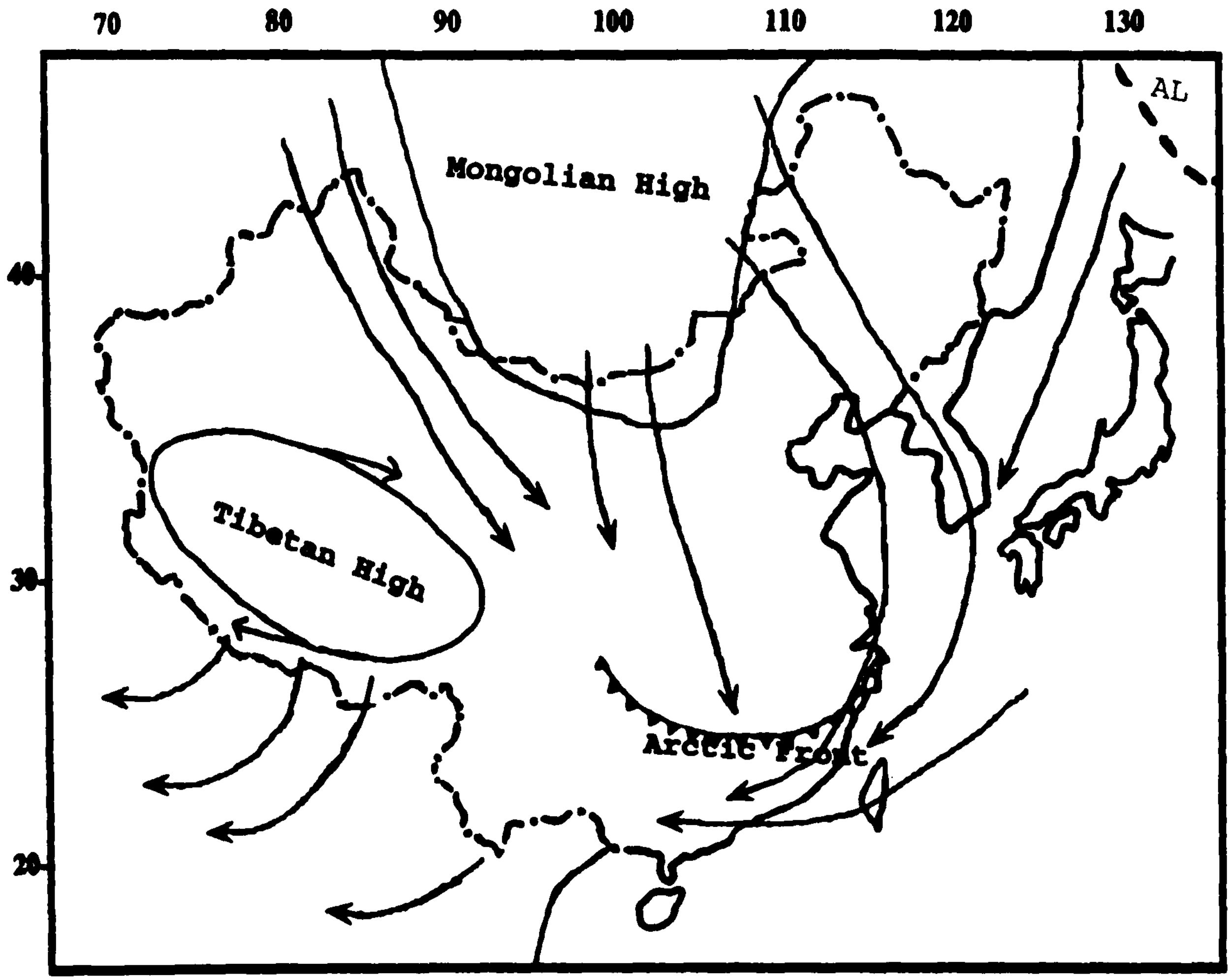


Figure 3.2 The winter circulation over China. (Modified from Zhang, 1984; Jarvis, 1993). AL represents the Aleutian Low.

are from N, NW, and NE in winter, while winds turn to east over the Qinghai-Tibetan Plateau (Domros and Peng, 1988; Zhang and Lin, 1992). Under this circulation, most of China is under a cold and dry climatic condition (Ren, 1981).

During the summer, the Qinghai-Tibetan Plateau warms rapidly relative to the Indian Ocean, resulting in low pressure over Asia and higher pressure over the ocean (Murakami, 1987; Overpeck et al., 1996). The strong low-level atmospheric pressure gradient in turn generates the SW and SE monsoon (Figure 3.3). Most of China is warm and wet.

The Qinghai-Tibetan Plateau shows a marked gradient in precipitation from east to west, related to the decreasing influence of the SW monsoon (Figure 3.4). Precipitation falls in summer as convective rain or snow (Tang et al. 1979; Thompson et al., 1989). Mean annual precipitation ranges from 25 mm to 600 mm on the Qinghai-Tibetan Plateau (Beijing Institute of Geography, 1990).

3.3 Vegetation

The geographical distribution of vegetation on the Qinghai-Tibetan Plateau is primarily controlled by climate (Zheng et al., 1979). In the southeastern part of the Qinghai-Tibetan Plateau where the annual precipitation reaches 600 mm, warm temperate deciduous broad leaved forest and alpine forest meadow have developed (Figure 3.5). On the vast flat of the central Qinghai-Tibetan Plateau, the annual

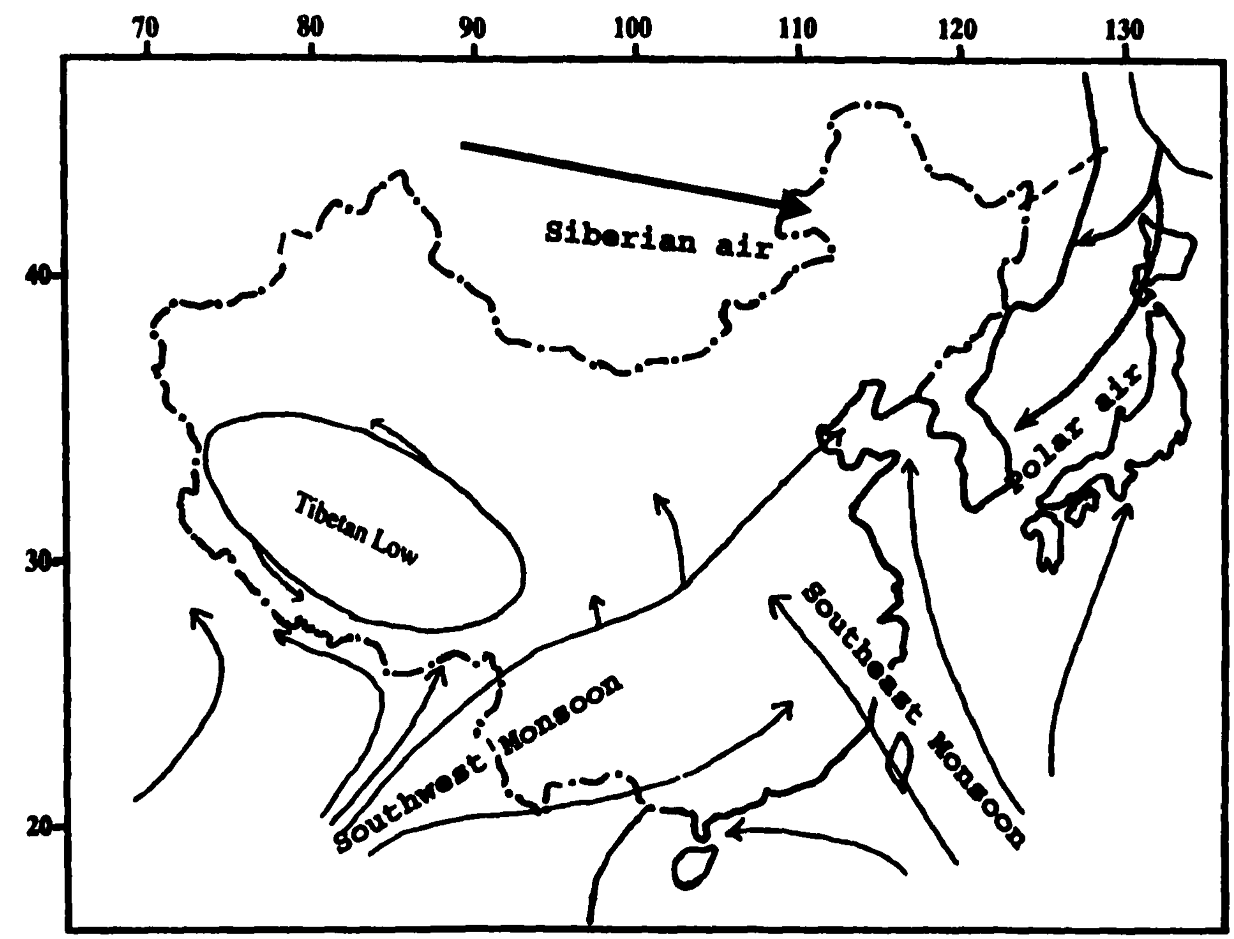


Figure 3.3 The summer circulation over China. (Modified from Zhang, 1984; Jarvis, 1993).

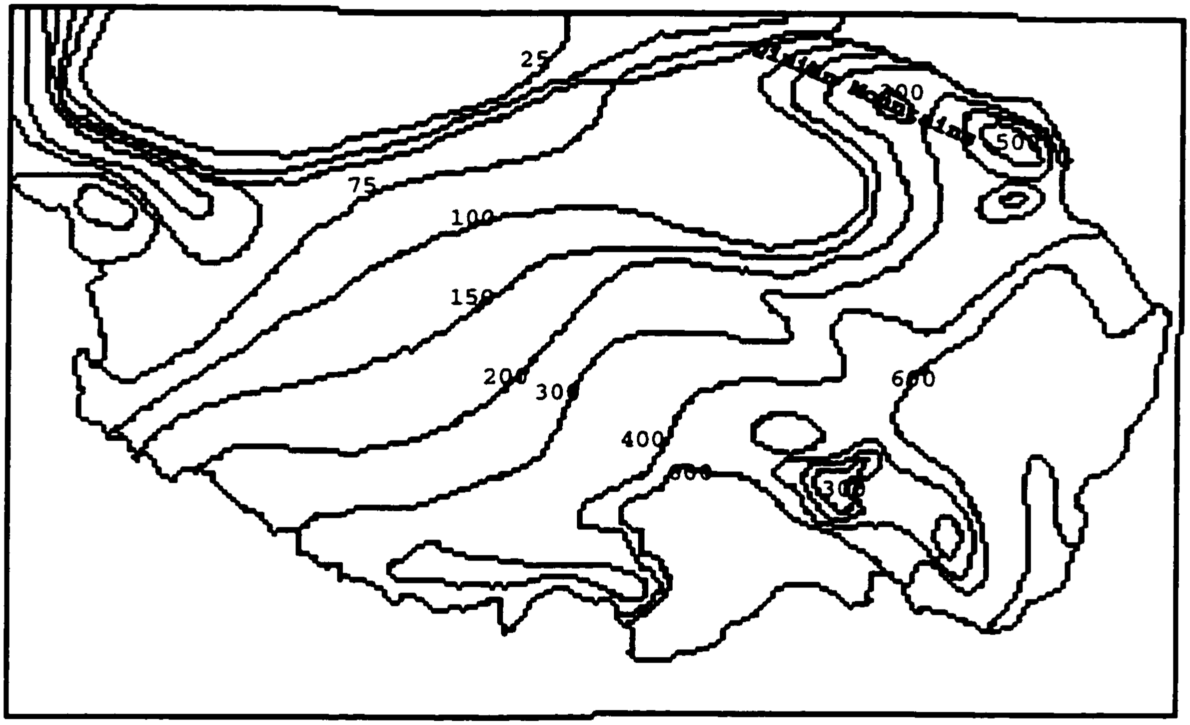


Figure 3.4 Map of annual precipitation (mm) over Qinghai-Tibetan Plateau. (After Beijing Institute of Geography, 1990).

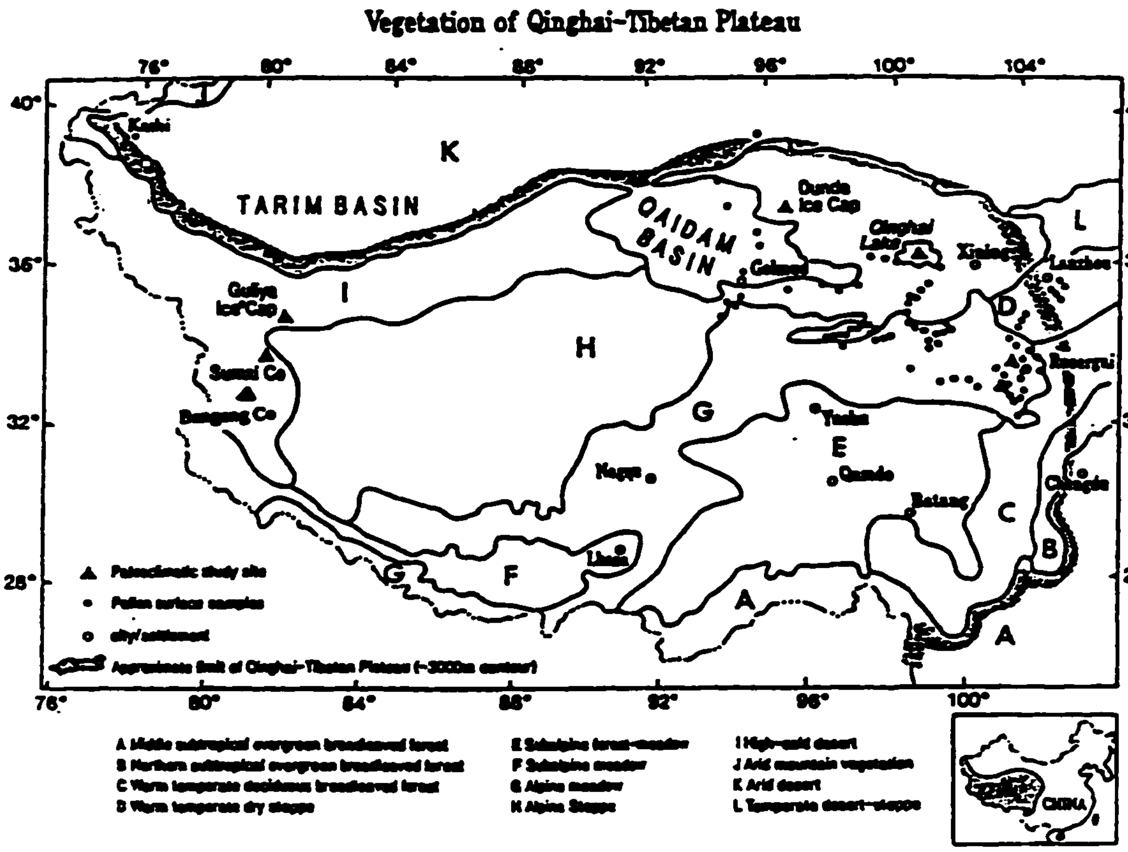


Figure 3.5 Map of vegetation regions of the Qinghai-Tibetan Plateau. (Modified from Jiao, 1984). Location of 87 pollen surface samples collected by Liu in the summer of 1993.

precipitation decreases to about 200 mm, and steppe vegetation prevails. Between the two vegetation zones, alpine meadow has developed. In the western and northwestern part, deserts occur because the climate is very dry and cold. Therefore, on the Qinghai-Tibetan Plateau, vegetation changes from southeast to northwest, grading from forest through alpine meadow, steppe, then to desert.

3.3.1 Alpine Forest Meadow

It occurs in the southeastern part of Qinghai-Tibetan Plateau. Here the elevations are generally below 3800 m and the precipitation is around 400-700 mm per year (Zheng et al, 1979). The vegetation is mostly dominated by *Betula*, *Quercus*, and *Picea*. The trees are tall, sometimes reaching 15-20 m. In dry sites and on south-facing slopes, *Sabina* may form the upper treeline at elevation of 4000 m or more.

3.3.2 Alpine Meadow

They are found today between 3000-4000 m and up to 5000 m (Zheng, 1996; Le Houerou, 1987). The zone of alpine meadow is located in the middle-eastern part of the Qinghai-Tibetan Plateau (Zheng, 1996). The climate of this zone is very cold, characterized by low temperatures in the warm season with a mean temperature of 6-12°C. Annual precipitation is about 250-600 mm (Zheng et al., 1979). The evaporation/ precipitation ratio is about 1.0-3.0. Westerlies prevail in winter. It is characterized by severe cold and drought with occasional snow in winter and spring.

The alpine meadows are characterized by a low growth layer, simplified structure with usually one herbaceous layer, dense plants, large coverage, short growing season and low production of biomass (Zheng, 1996). The most important taxon forming the alpine meadows is *Kobresia*, which may be mixed with other sedges and grasses such as *Carex*, *Festuca*, and *Poa*. Other common herbs include *Stellera*, *Potentilla*, *Ranunculus*, *Anemone*, *Polygonum*, and *Aster* (Zheng et al, 1979; Zheng, 1996). In wetter places or where drainage is impeded, swampy meadows occur. These occur at higher elevations above 4400 m. The ground cover in the alpine meadows may be thin and cushion plants such as *Androsace* are common. On the more exposed slopes, or where overgrazing occurs, the alpine meadow may have a more degenerative appearance with lower coverage (Zheng, 1996).

Montane coniferous forests occur in patches only in the peripheries of the region (Wu, 1980; Zheng, 1996). Forests of *Picea balfouriana* appear on shady slopes at an elevation of 3400-4000 m asl. Forests of *Picea crassifolia* can be found at elevation of below 3500 m asl. On the eastern flanks of Anyemaqen Mountains, open forests of *Sabinia tibetica*, *S. convallium*, and *S. przewalskii* exist on the sunny slopes (Zheng, 1996).

3.3.3 Steppe

It is widespread in central and northeastern Qinghai-Tibetan Plateau at elevations below 3500 m, but it can also occur on wind-swept high plains above 4200 m. The warmest monthly temperature is 9-12°C. The annual precipitation is about 150-400 mm (Table 3.1). According to the ecological-physiognomic vegetation classification of Wu (1980), the steppes of the Qinghai-Tibetan Plateau can be divided into two subtypes: high-cold steppe and montane steppe. The high-cold steppe is dominated by perennial grasses and low shrubs adapted to cold and drought. The dominants in the montane steppe are commonly perennial grasses that are adapted to the temperate or warm-temperate and dry climate. Table 3.1 shows the climatic characteristics of steppes.

Poaeaea, Asteraceae, Fabaceae and Cyperaceae are most important families in the steppe vegetation of the Qinghai-Tibetan Plateau. Important genera are *Artemisia*, *Carex*, *Stipa*, *Orinus*, and *Pennisetum* (Wang, 1981, 1988). *Stipa* and *Artemisia* are the most important plants in the steppes (Wang 1981). Some *Picea crassifolia* and *Sabina przewalskii* are scattered on the northeastern Plateau. In drier sites these

Table 3.1 Climatic characteristics of steppe areas of the Qinghai-Tibetan Plateau. (Modifed from Wang, 1988).

	Site	Altitude (m)	Ann. mean (°C)	Mean temp. warmest month (°C)	Ann.mean rainfall (mm)	Relative humidity (%)	Evaporative capacity (mm)
High-cold Steppe	Pangkog Co	4380	-2.2	10.3	246	47	2239
	Xainza	4670	-0.3	9.4	291	40	2167
	Gerze	4415	0.1	12.1	166	34	2418
Montane Steppe	Xigaze	3836	6.3	14.6	434	41	2441
	Gyangze	4040	4.7	12.8	288	41	2569
	Burang	3900	3.0	13.7	169	46	2181

trees may co-dominate with *Artemisia*. Other common grasses include *Achnatherum* and *Poa*.

3.3.4 Desert

Desert occurs in the western part of the Qinghai-Tibetan Plateau. The desert in the Qinghai-Tibetan Plateau covers a much smaller area than that of the steppe. In the desert region, the warmest monthly temperature is 4-16°C (Wang, 1988). There is very little precipitation less than 50 mm annually. The evaporation/precipitation ratio is 6.1-15.0 (Wang, 1988). Strong wind occurs during winter and spring. A sparse cushion-like plant cover exists here (Wang, 1988; Zheng et al., 1979).

The dominants of the desert are mainly Chenopodiaceae and Asteraceae (Zheng et al, 1979; Wang, 1988). Other important families are Poaceae, Fabaceae, Brassicaceae, Boraginaceae, and Ephedraceae. According to the ecological-physiognomic vegetation classification (Wu, 1980), the desert of the Qinghai-Tibetan Plateau can be divided into two subtypes: high-cold desert and montane desert.

The dominants of high-cold desert are cushion-like minor semi-shrubs adapted to extremely cold and dry climates. The representative community is dominated by *Ceratoides compacta* (Wang, 1988). The dominants of montane desert are the xerophilous minor semi-shrub. The representative communities are dominated by *Ceratoides latens* and *Ajania fruticulosa*.

3.4. The Ice Cores

3.4.1 The Dunde Ice Cores

The Dunde Ice Cap (38°06′N, 96°24′E, summit 5325 m a.s.l) is located in a desert environment between the highest Chinese desert, the Qaidam Basin, to the south and west, and the Gobi Desert to the north (Figure 3.6). This relatively large ice cap has a summit elevation of 5325 m and a total area of 60 km^2. The Dunde area is characterized by low temperature. Mean annual temperature at the Dunde Ice Cap is about -7.3°C (Thompson et al., 1989). During the three-month growing period, monthly mean temperatures between June and September are less than 15°C.

In 1986 and 1987, two shallow ice cores over 30 m in length and three complete ice cores around 140 m in length were recovered from the Dunde Ice Cap summit and brought back to the United States (Thompson et al., 1989) (Figure 3.6). In the field, Deep Core D-1 was cut into 3585 samples that were melted in closed containers by passive solar heating in a laboratory tent. The upper 56 m of core D-3 were melted and bottled on site while the lower 83 m were kept frozen. Core D-2 was kept in China.

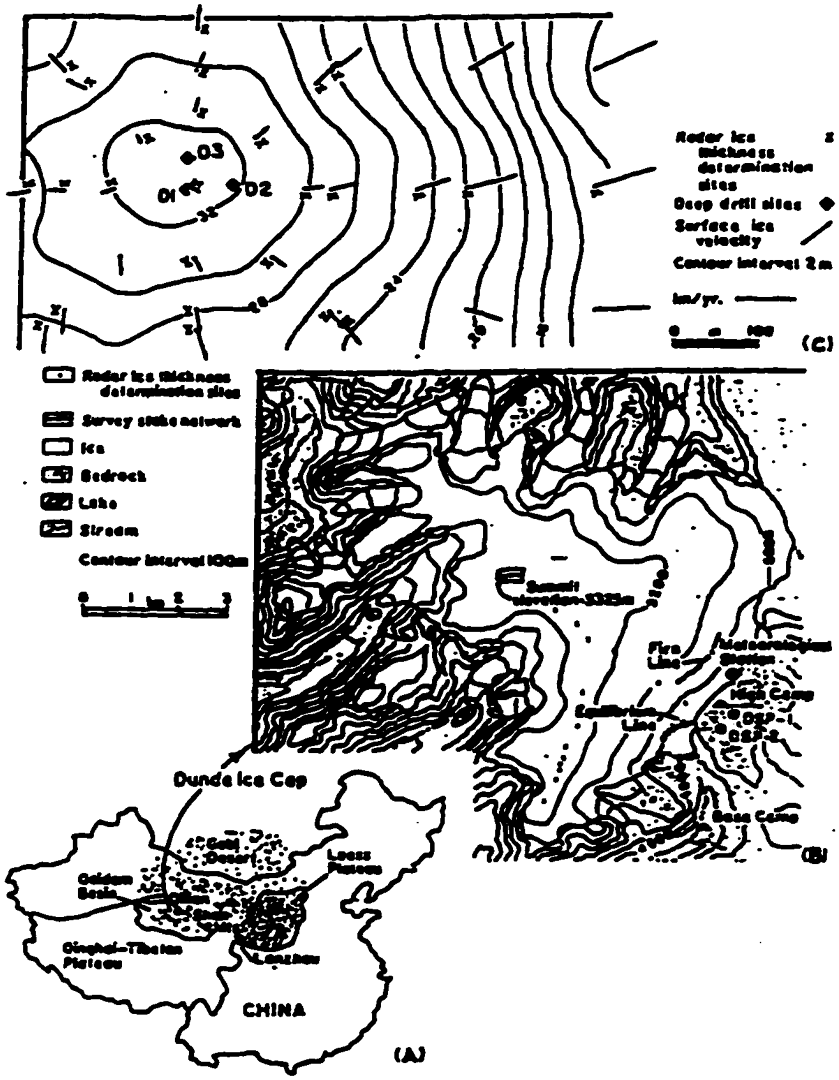


Figure 3.6 (A) Location of the Dunde ice cap in relation to the Qinghai-Tibetan Plateau, Qaidam Basin, Gobi Desert, and Loess Plateau; (B) Topography of the Dunde ice cap; (C) Location of the core sites, D-1, D-2, and D-3. (After Thompson et al., 1989).

Figure 3.7 shows the relationship between depth and age in the Dunde ice cores. Dating of the ice core was accomplished with the use of several techniques (Thompson et al., 1989; Yao and Thompson, 1992). The time model since 1400 has been established according to the seasonal variations of $\delta^{18}O$, microparticle concentrations, and electrical conductivity (Figure 3.8). Counting the annual $\delta^{18}O$ and microparticle concentration peaks to 70 m and the visible dust layers below 70 m indicated that the ice at 117 m depth was deposited about 4550 years ago (Thompson et al., 1989). The age of the deeper ice can be roughly estimated by extrapolation of the annual layer thickness data. At a depth of 129.2 m in the D-1 core, the age is then calculated to be about 11950 years ago. In consideration of the uncertainty in the accumulation rate assumption, this age is close to the annually dated age of 10750 years ago for the LGS-Holocene transition in the Camp Century core from northern Greenland (Thompson et al., 1989). At the bottom, ice was estimated to be more than 100,000 years old by using the ice flow model.

3.4.2 The Guliya Ice Cores

The Guliya Ice Cap (35°17'N, 81°29'E; summit 6710 m a.s.l.) is in the far western Kunlun mountains on the Qinghai-Tibetan Plateau, China (Figure 3.9). It is the highest, largest (>200 km^2) and thickest (308.6 m)

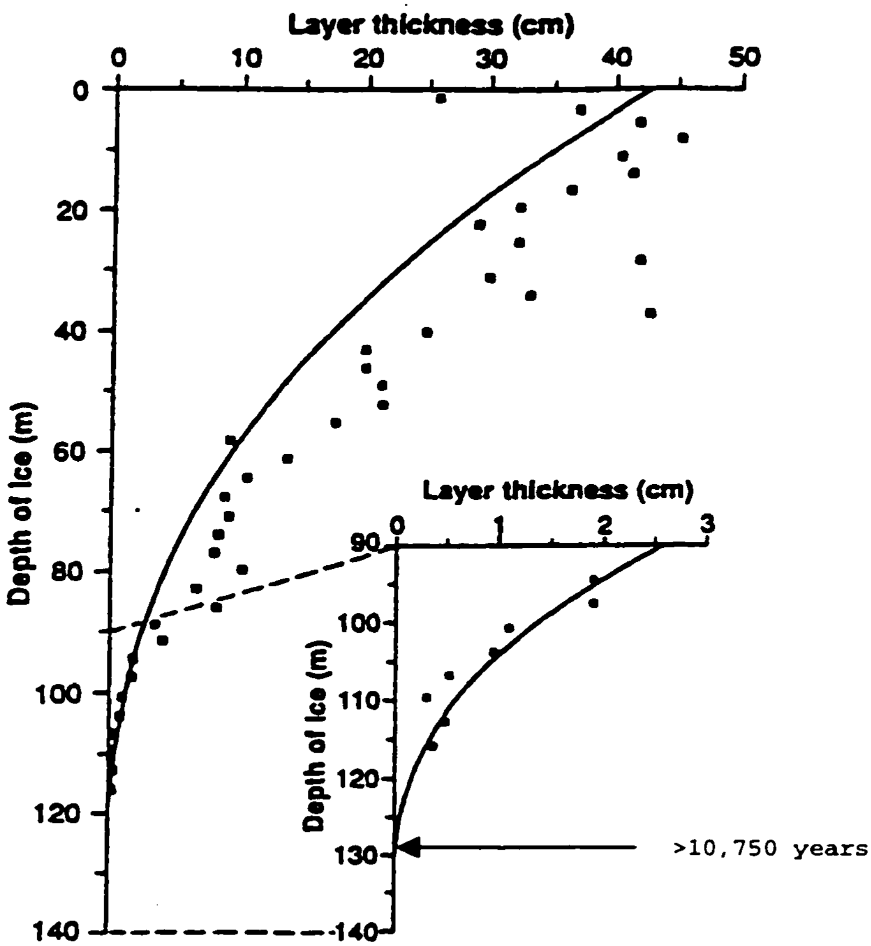


Figure 3.7 Average annual layer thickness (black squares) based on the visible dust layers plotted as a function of depth in core D-1. The inset shows the detailed thickness below 90 m. (After Thompson et al., 1989)

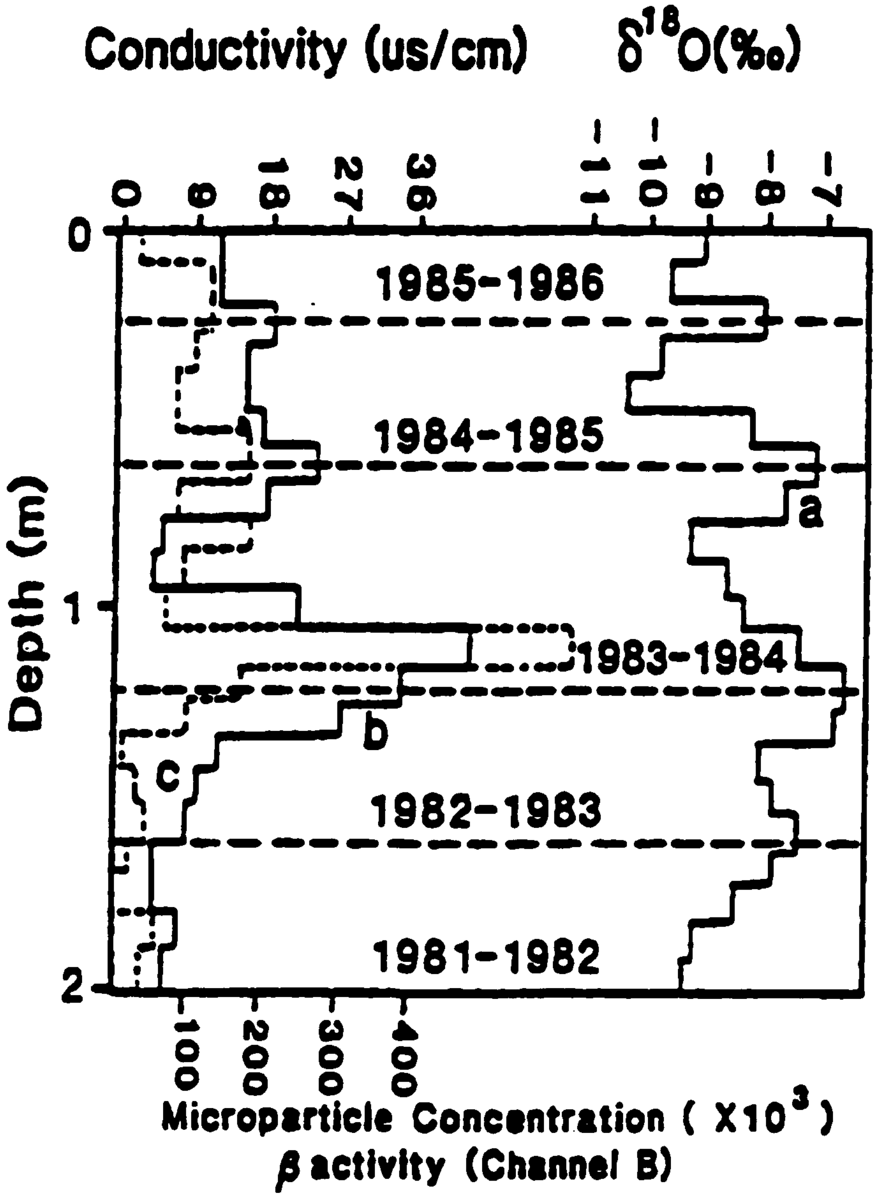


Figure 3.8 An example of dating the Dunde ice core by combining oxygen isotope, microparticle conductivity, and β activity data. Line a is $\delta^{18}O$, b is microparticle concentration, and c is conductivity. (After Yao and Thompson, 1992).

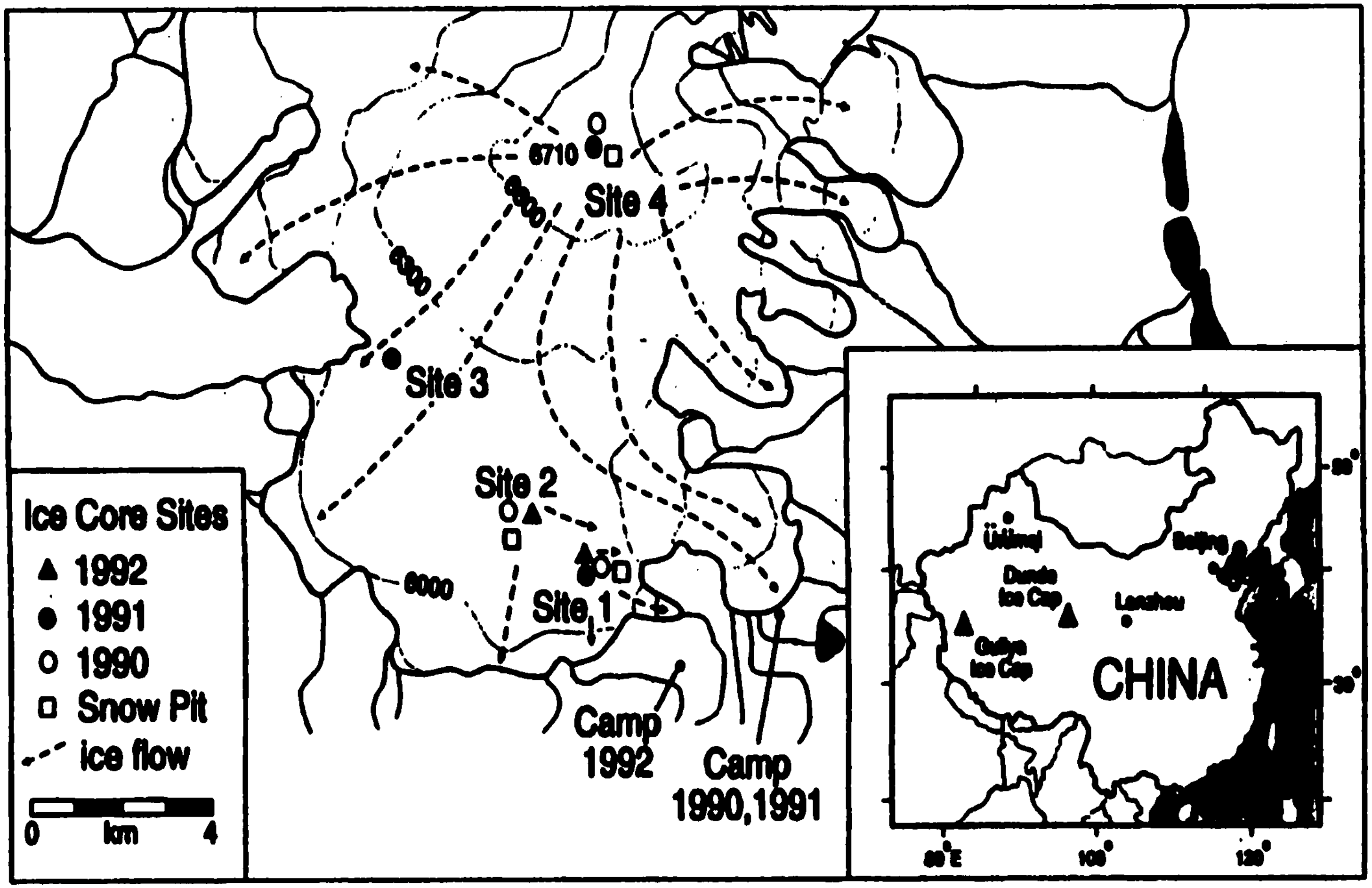


Figure 3.9 Locations of of snow-pit and coring sites on the Guliya ice cap. (After Thompson et al., 1995a).

subtropical ice cap yet investigated (Thompson et al., 1995a). In 1992, an American-Chinese expedition successfully recovered a 308.6-meter ice core from the Guliya Ice Cap. The Guliya Ice Cap is surrounded by vertical 30 to 40 meter ice walls. It resembles a "polar" ice cap and has internal temperatures of -15.6°C, -5.9°C, and -2.1°C at 10 m, 200 m and the base, respectively.

The 308.6 m core to bedrock provides the first high resolution ice-core record of the last glacial cycle as recorded in the subtropics. The entire length of the frozen core was analyzed by cutting 12,628 samples for oxygen isotopic ($\delta^{18}O$) measurements, 12,480 samples for dust concentrations, and 9,681 samples for anion Cl^-, NO^{3-}, and SO_4^{2-} concentrations (Thompson et al., 1997).

Figure 3.10 shows the depth-age model of the Guliya ice core. The time-scale from the Guliya ice cores comes from several dating techniques. The upper 120 meters is dated by counting the annual ice and dust years. Below 120 meters, dating is based on the apparent correlation between atmospheric CH_4 levels and stadial and interstadial events inferred from $\delta^{18}O$ values in polar ice cores to establish the Guliya time scale for the past 110,000 years (Thompson et al., 1997). The Guliya $\delta^{18}O$ record is linked by extrapolation

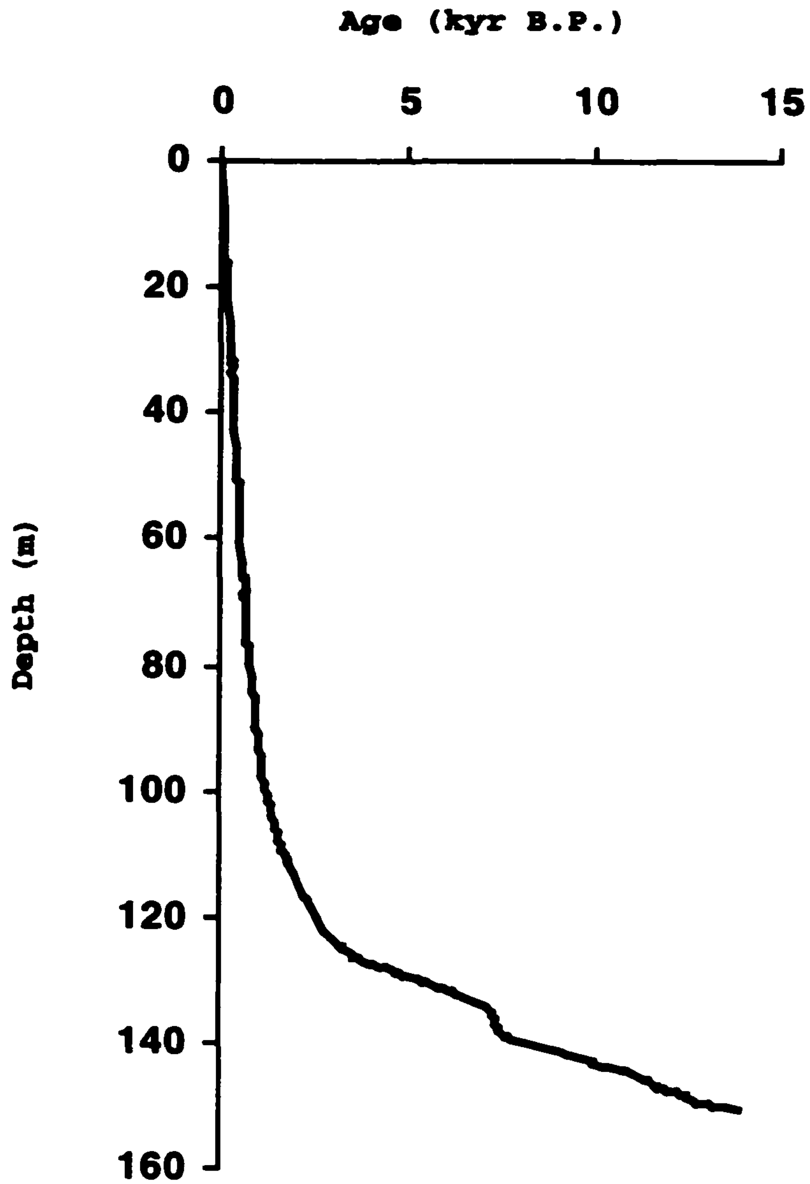


Figure 3.10 The relationship between depth and age on the Guliya ice core. (Data from L.G. Thompson, the Ohio State University).

to the SPECMAP chronology, under the assumption that the methane fluctuations in high-latitude ice cores are synchronous with stable isotope variations in western China (Thompson et al., 1997). Using ^{36}Cl dating method, the ice below a depth of 290 m is estimated to be 500,000 years old (Thompson et al., 1997).

3.5 Materials and Methods

3.5.1 Materials

For pollen study, meltwater samples were used from both the Guliya and Dunde ice cores. Samples were shipped from the Byrd Polar Research Center, Ohio State University, to Louisiana State University. All meltwater samples are stored in plastic bottles.

3.5.2 Pollen Sampling

The Dunde Ice Core

For Dunde, meltwater samples come from two long cores, D-1 and D-3, and were supplemented by samples from shallow cores 1 and 2. There were over 2000 bottles of meltwater samples in total. The capacity of a bottle is 150 ml. In order to get enough pollen four or more bottles of meltwater were combined for each pollen sample.

The sampling resolution for pollen study decreases downward from the upper part of the ice core due to the exponential thinning of the annual layers downcore. The seasonal pollen samples are distinguished by using seasonal variations of oxygen isotope, dust, and electrical conductivity. Less negative $\delta^{18}O$ values, low values of dust, and electrical conductivity represent summer season, while more negative $\delta^{18}O$ values and higher value of dust and electrical conductivity represent winter. Finally, there are 120 seasonal pollen samples covering the interval from A.D. 1927 to A.D. 1986. The seasonal samples come from three different cores: the 1977-1986 samples from shallow core 2, the 1947-1986 samples from deep core 3, and the 1927-1946 samples from shallow core 1.

Beyond 1927, each pollen samples is a composite of 4 bottles of meltwater. Based on the depth-age model provided by Thompson et al. (1989), the pollen samples ranged from one year to 1000 years. Time resolution decreases per pollen sample: seasonal for the last 60 years (i.e. 1927 to 1986), annual for the last 100 years, decadal at about 230 BP (calendar years before 1986), centennial at about 2300 year BP, and millennial at the Holocene/Late Pleistocene transition at about 11,000 BP. Altogether more than 560 samples were analyzed for pollen in the Dunde ice cores.

The Guliya Ice Core

In the Guliya ice core, meltwater samples stored in 250 ml bottles were shipped from the Ohio State University to Louisiana State University. Forty-one (41) meltwater samples were used for pollen analysis from long cores D-1. These meltwater samples ranged in volume from 250 to 850 ml (mean = 650 ml). From present to 9000 BP, each sample represents a 50-year aggregate, and the interval between adjacent samples is 500 years. Beyond 9000 BP, each sample represents about 100 years.

3.5.3 Pollen Extraction

Meltwater samples were put on a hot plate to evaporate (without boiling) to reduce the volume of water to less than 100 ml, then centrifuged. All samples were treated with acetolysis solution. No HF was used in the procedure so that observation of the dust characteristics would be possible. A known quantity of marker *Lycopodium* spores was added to each sample before processing to determine pollen concentration. All pollen samples were mounted in silicon oil and examined at magnifications of 400X and 1000X under the microscope.

3.5.4. Pollen Counting

My identifications of pollen were based on comparison with a modern pollen reference collection of Dr. Kam-biu Liu at the Louisiana State University, and with published Asian pollen atlases and keys (Institute of Botany, 1960, 1976; Huang, 1972; Institute of Botany and South China Institute, 1982). At least 1000 *Lycopodium* markers were counted per sample.

3.6. Statistical Methods

3.6.1 Cluster Analysis

Cluster analysis is specifically a classificatory technique, designed to select subsets of mutually similar objects from the set of all such objects (Mather and Doornkmap, 1970). This aim may be achieved in several ways (Norusis, 1988). The SPSS computer program (version 7) for cluster analysis was used. In this study, the Ward's method (Squared Euclidean Distances) was used because it produced the most meaningful results. Seventeen pollen types each with a mean percentage of >1.0 are used as variables.

3.6.2 Discriminant Analysis

Discriminant analysis identifies that linear combination (or combinations) of a set of independent variables which best discriminates among a number of previously defined groups (Liu and Lam, 1985). The principles and applications of discriminant analysis in palynological research are described by Liu and Lam (1985). In this study, discriminant analysis is used to test the validity of classification of surface samples.

3.6.3 Trend-surface Analysis

Local anomalies in pollen distribution may hinder interpretation of regional distribution patterns. For any geographically distributed set of data, trend-surface analysis seeks to filter out local differences among samples and to emphasize the regional pattern. This technique fits a polynomial function of the geographic coordinates (x for latitude and y for longitude) to the percentages of the ith pollen type (Pi) and thus approximates the geographic trend of that pollen distribution by an nth-degree equation of the form

$$Pi=a_1X^n+a_2Y^n+a_3X^{n-1}Y+\ldots+a_{q-1}y+a_q,$$

Where

$$Q = \sum_{k=1}^{N+1} k$$

and the a_j ($j = 1, \ldots, q$) values are regression coefficients. The derived equation has predictive value and can be used to estimate pollen percentages at a station given its latitude and longitude. For this study, three degree trend surfaces ($n = 3$) were calculated. The surfaces are sufficiently general to filter out the effects of individual sites and present a summary of the regional patterns in the data.

CHAPTER IV

MODERN POLLEN RAIN STUDY

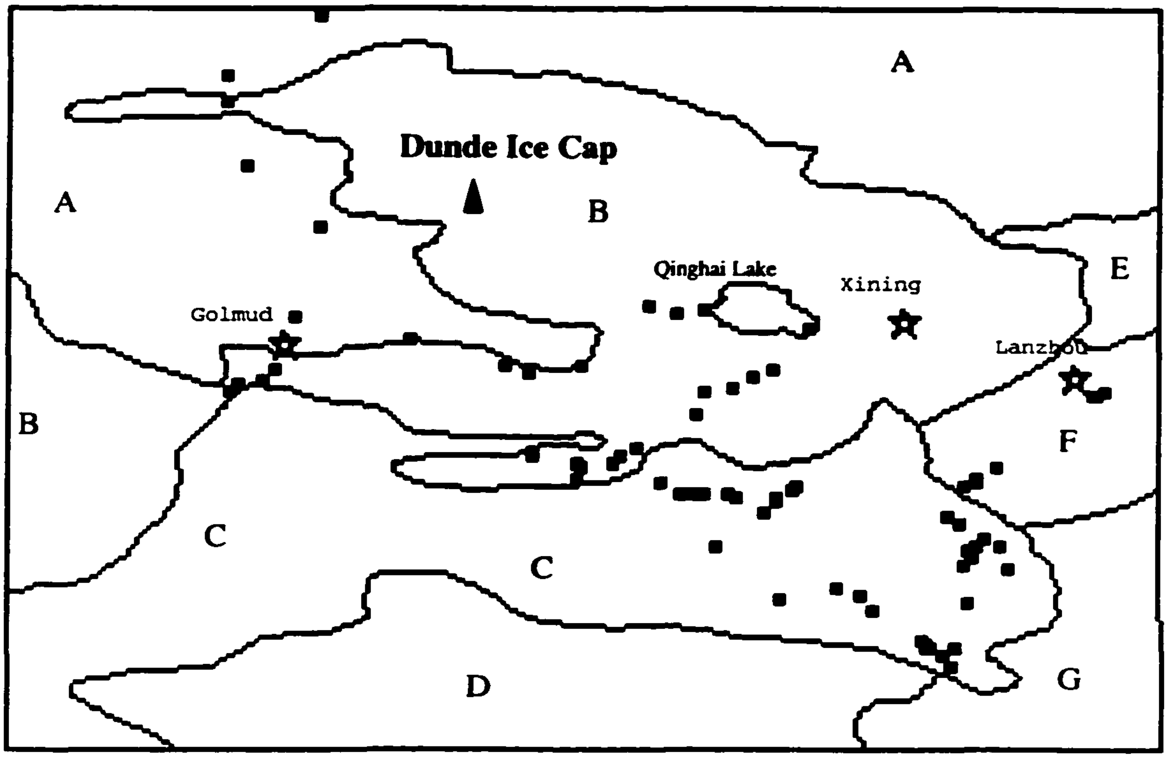
4.1 Introduction

Paleoecologists using pollen analysis to reconstruct changes in past vegetation need first to determine how the modern vegetation is represented by its pollen rain (Webb and McAndrews, 1976; Webb et al., 1987, 1993). In the Qinghai-Tibetan Plateau, few studies of the relationship between vegetation and pollen rain have been conducted (Van Campo et al., 1996).

In order to establish the relationship between pollen and vegetation, in the summer of 1993 Kam-biu Liu (Louisiana State University) and two Chinese scientists (Liu Guangxiu and Chen Guichen) from the Lanzhou Institute of Glaciology and Geocryologist and the the Qinghai Institute of Plateau Biology, respectively, undertook a month-long expedition to collect surface samples from the northeastern Qinghai-Tibetan Plateau. During this expedition, eighty-seven (87) pollen surface samples were collected from regions southeast to northwest of the Dunde Ice Cap (Figure 4.1).

4.2 Source of the Data

Most of the surface samples were collected from moss growing on soil surface in the open landscape of the Qinghai-Tibetan Plateau. Several (mainly from desert) are



A Arid desert B Alpine steppe C Alpine meadow
D Subalpine forest-meadow E Warm temperate dry steppe F Temperate desert steppe G Warm temperate deciduous forest.

Figure 4.1 Location of the 87 surface samples in relation to the regional pattern (after Jiao, 1984) of the Qinghai-Tibetan Plateau. The stars represent city, dots represent sample location.

top-soil samples collected from microtopographic depressions sheltered from strong wind and where water and finer sediments collect after rain. Some samples were collected from shallow pools or the edge of small natural ponds (Liu, 1997). More than 300 pollen grains were counted for each sample. Fifty-four (54) pollen types were identified.

4.3 Vegetation

The surface samples are located broadly along a southeast to northwest transect that straddles four major vegetation regions: warm temperate dry steppe; alpine meadow; alpine steppe, and desert (Zheng et al, 1981; Jiao, 1984). At a local scale, five vegetation communities were recognized in the field---forest meadow; shrub meadow; alpine meadow; steppe; and desert. These vegetation communities are briefly described as follows.

4.3.1 Forest Meadow

These forests usually have an open canopy, and are surrounded by or intermingled with alpine meadows. They are mostly dominated by *Picea*, together with some *Betula*. The trees are tall, sometimes reaching 15-20 m. In dry sites and on south-facing slopes, *Sabina* may form the upper treeline at elevations of 4000 m or more (Liu, 1997).

4.3.2 Shrub Meadow

This is a very common vegetation community that occurs locally in the alpine meadow region at elevations of >3500 m. They are typically found on north-facing slopes or in relatively moist sites, and merge gradually with the surrounding alpine meadow (Liu, 1997). The most common shrub taxa in the shrubs meadows are *Potentilla*, *Spirare*, *Salix*, *Rhododendron*, and *Caragana* (Liu, 1997).

4.3.3 Alpine Meadow

The most important taxon forming the alpine meadows is *Kobresia*, which may be mixed with other sedges and grasses such as *Carex*, *Festuca*, and *Poa*. Other common herbs include *Stellera*, *Potentilla*, *Ranunculus*, Anemone, *Polygonum*, and *Aster*. On the more exposed slopes, or where overgrazing occurs, the alpine meadow may have a more degenerative appearance with lower coverage (Liu, 1997).

4.3.4 Alpine Steppe

This community is widespread in areas south and southwest of Qinghai Lake at elevations below 3500 m, but it can also occur on wind-swept high plains above 4200 m. *Stipa* is the most important grass in the alpine steppes (Wang, 1981). In dry sites it may co-dominate with *Artemisia*. Other common grasses include *Achnatherum* and *Poa* (Liu, 1997).

4.3.5 Alpine Desert

Shrub desert occurs in the eastern part of the Qaidam Basin. Common shrubs are *Nirtraria, Kalidium, Salsola, Ceratoides, Calligonum,* and *Ephedra*. In the central and western part of the Qaidam Basin where the annual precipitation is <50 mm, the landscape is typically a Gobi Desert virtually devoid of vegetation cover. Eolian landforms are common, such as branches and deflation basins. Salt lakes occur because of the very low precipitation and strong evaporation (Chang, 1981, 1983; Liu, 1997).

4.4. Results

4.4.1 Cluster Analysis

Using Ward's method in cluster analysis (SPSS software 7.0), the 87 samples were grouped into four groups (Figure. 4.2). These four groups correspond closely with the five local vegetation communities (desert, alpine meadow, alpine shrub, alpine steppe, and forest-meadow) that these samples were identified with in the field. Samples from alpine meadow and alpine shrubs both are characterized by high percentages of Cyperaceae (Figure 4.3). It is not surprising that the palynological signatures of these two vegetation types are indistinct from each other, since the alpine meadow and alpine shrub communities tend to occur side-by-side (Figure 4.4). The samples from the forest meadow are characterized by high percentages of *Picea*, as well as some

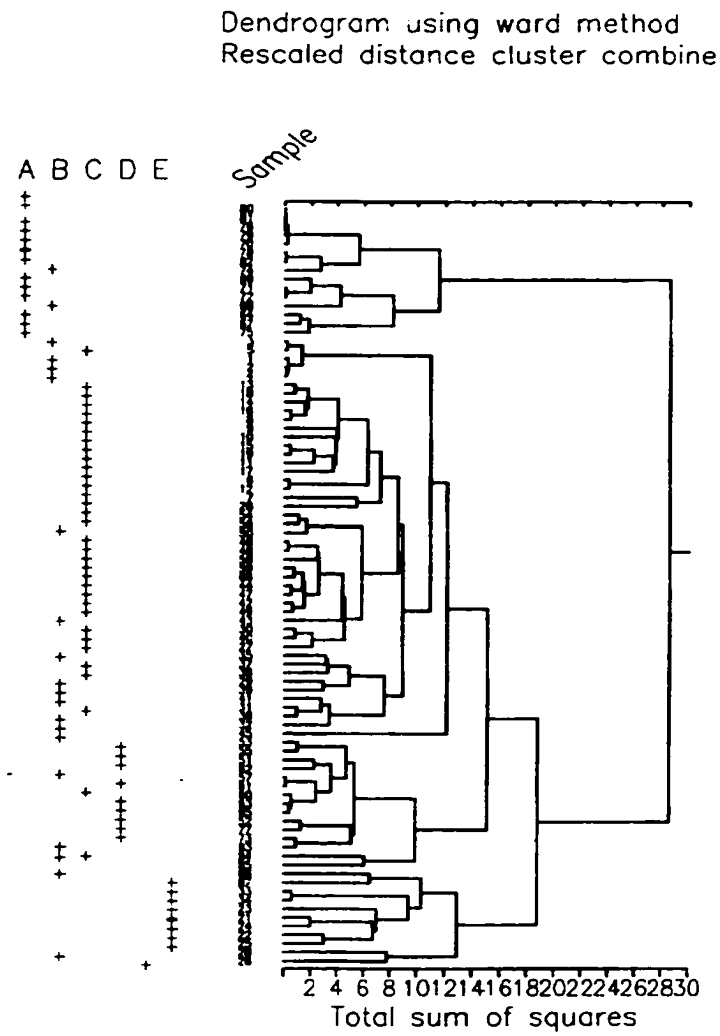


Figure 4.2 Classification of the 87 surface samples by cluster analysis. The columns on the left denote the original classification of the samples in the field according to their local vegetation communities. A=desert; B= alpine meadow; C= shrub meadow; D=steppe; E=forest meadow. The five clusters classified by cluster analysis correspond broadly with five local vegetation types identified in the field.

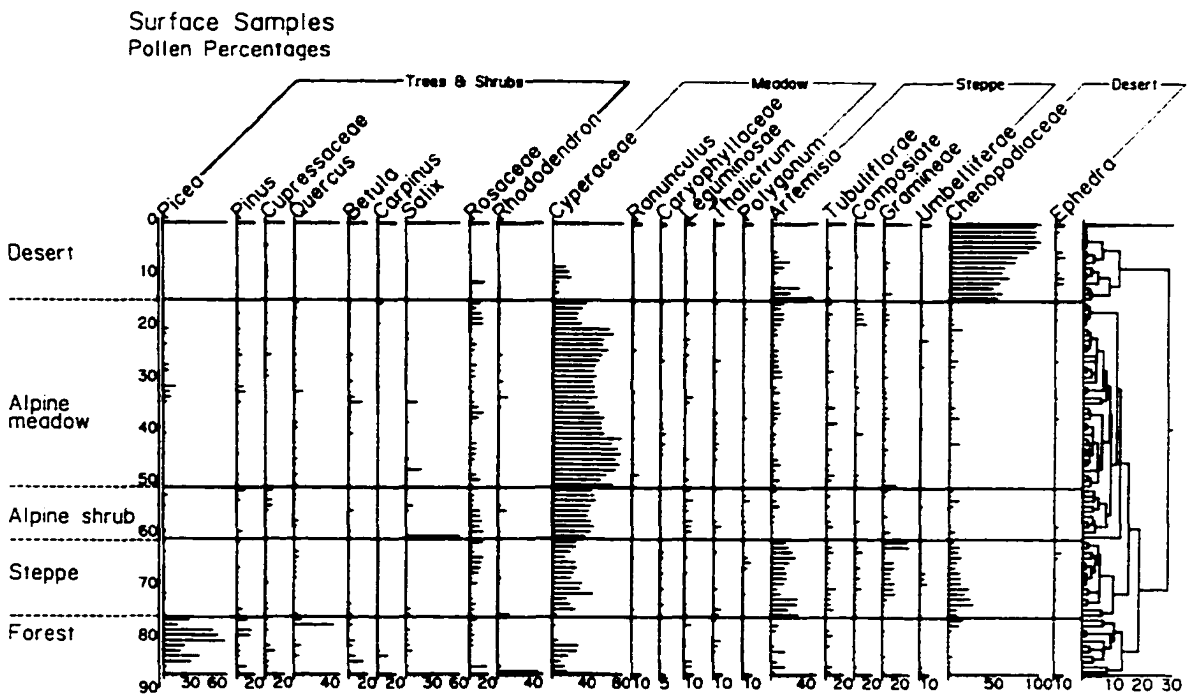


Figure 4.3 Pollen percentage diagram of the 87 surface samples collected from the northeastern Qinghai-Tibetan Plateau. The samples are arranged by vegetation types in accordance with the five groups delineated by cluster analysis using Ward Method. The pollen sum includes all pollen and spore taxa.

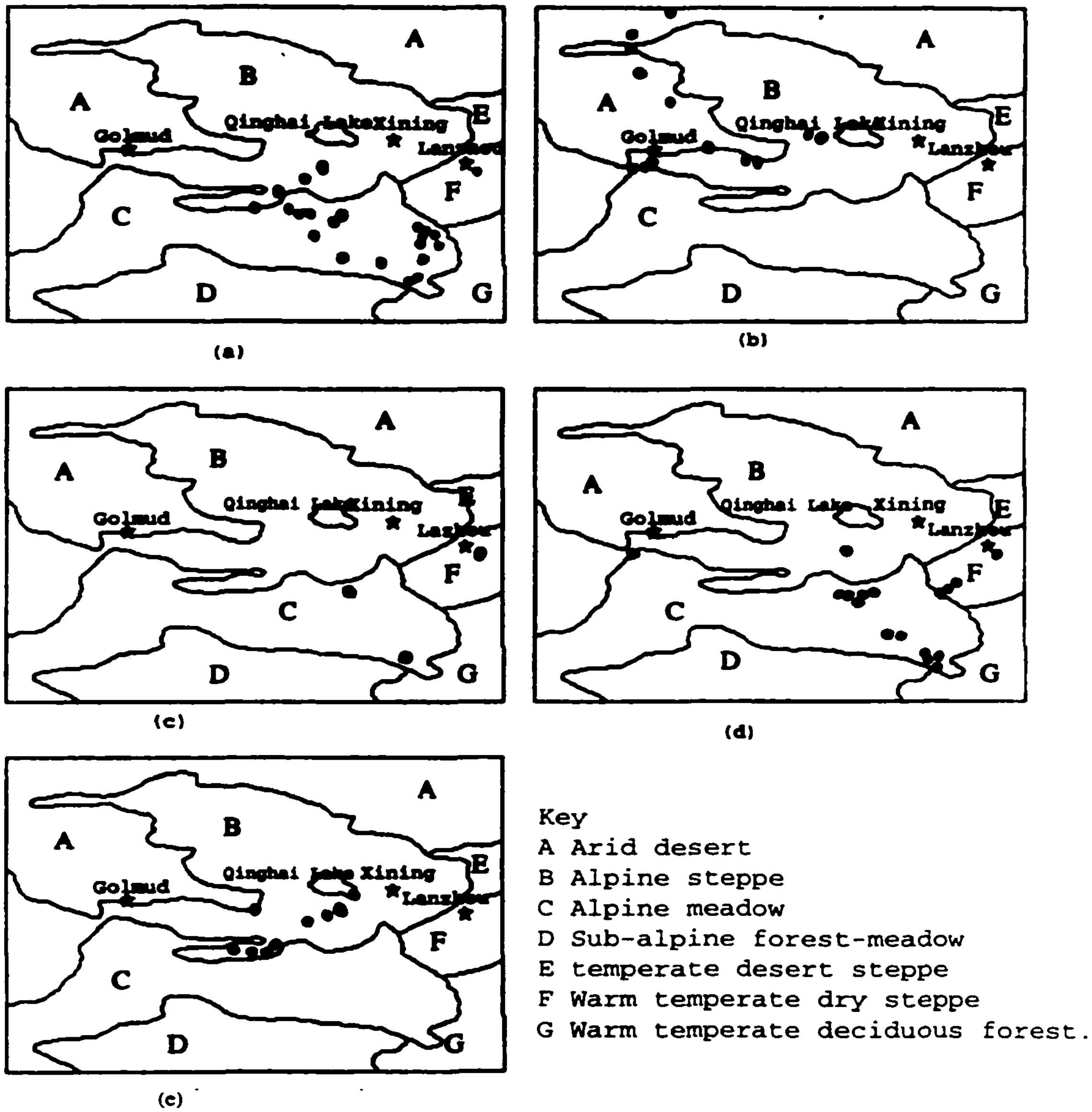


Figure 4.4 Relationship between surface samples and vegetation regions. Dots represent the locations of samples. (a) alpine meadow samples related to vegetation; (b) desert samples related to vegetation; (c) alpine forest samples related to vegetation; (d) alpine shrub samples related to vegetation; (e) alpine steppe samples related to vegetation.

Pinus and *Betula*. The dominant pollen types are similar to the predominant plant communities in forest meadow. The steppe samples are characterized by consistently high percentages of *Artemisia* (30%), Cyperaceae (25%), Chenopodiaceae (20%) and Gramineae (15%) pollen. In the desert group, Chenopodiaceae pollen is clearly the dominant, usually occurring at >50%.

4.4.2 Discriminant Analysis

Discriminant analysis (Liu and Lam, 1985) is used to evaluate objectively the classification of surface samples into groups corresponding to plant communities recognized in the field. The results show that 80.46% of the samples are correctly classified into the five vegetation groups (Table 4.1). Among the five groups, the forest meadow and desert have 100% and 85.7% of their samples correctly classified by discriminant analysis, respectively. Alpine meadow (79.3%), shrub meadow (75%), and steppe (75%) are less palynologically distinct, and each has a few samples misclassified into other groups.

Figure 4.4 shows spatial distribution of each sample related to vegetation. Six samples belonging to alpine meadow group lie in steppe vegetation area, and most of samples classified into shrubs are located in alpine meadow vegetation zone. Therefore, it is not surprising that there are some misclassification in alpine meadow and alpine

Tables 4.1. Classification results based on five groups classified in the field.

		No. of Predicted Group Membership				
Actual Group	Cases	Alpine meadow	Alpine shrubs	Forest-meadow	Steppe	Desert
Alpine meadow	29	23	4	0	2	0
		79.3%	13.8%	0%	6.9%	0%
Alpine shrubs	24	5	18	0	1	0
		20.8%	75.0%	0%	4.2%	0%
Forest-meadow	8	0	0	8	0	0
		0%	0%	100%	0%	0%
Steppe	12	2	0	0	9	1
		16.7%	0%	0%	75%	8.3%
Desert	14	0	0	0	2	12
		0%	0%	0%	14.3%	85.7.%

Percent of "grouped" cases correctly classified: 80.46%

shrubs. The classification success can be improved by combining the shrub meadow samples with the alpine meadow samples. Thus when four predetermined groups (forest meadow; alpine meadow-shrub meadow; steppe; arid desert) were used in the discriminant analysis, 89.66% of the samples are correctly classified (Table 4.2). By combining the alpine meadow and shrub meadow samples into one group (hear named alpine meadow), the percentage of correct classification increases to 90.6%. However, for the steppe group, the percentage is still low, because most of samples from this group are situated close to desert and alpine meadow (Figure 4.4) and contain greater amounts of locally dispersed pollen. Some samples located around Qinghai Lake (alpine steppe) contain high percentages of Chenopodiaceae pollen because of saline soil.

4.4.3 Trend-surface Analysis

Artemisia and Chenopodiaceae are two common pollen types in the Dunde area. The ratio between *Artemisia* and Chenopodiaceae pollen (A/C) has been used as an indicator of moisture availability in arid and semi-arid regions (Van Campo and Gasse, 1993; K-b Liu et al., 1998). The A/C ratio for each surface sample site was calculated. The A/C ratio isoline were mapped by means of trend surface analysis to highlight the general trends of spatial variations (Figure 4.5). The A/C ratio is between 5 and 10 in alpine meadow

Tables 4.2 Classification results based on four groups after combining the alpine meadow and alpine shrubs.

Actual Group	No. of Cases	Predicted Group Membership: Alpine meadow-shrub	Forest-meadow	Steppe	Desert
Alpine meadow	53	48	0	5	0
		90.6%	0%	9.4%	0%
Forest-meadow	8	0	8	0	0
		0%	100.0%	0%	0%
Steppe	12	2	0	9	1
		16.7%	0%	75.0%	8.3%
Desert	14	0	0	1	13
		0%	0%	7.1%	92.9%

Percent of "grouped" cases correctly classified: 89.66%

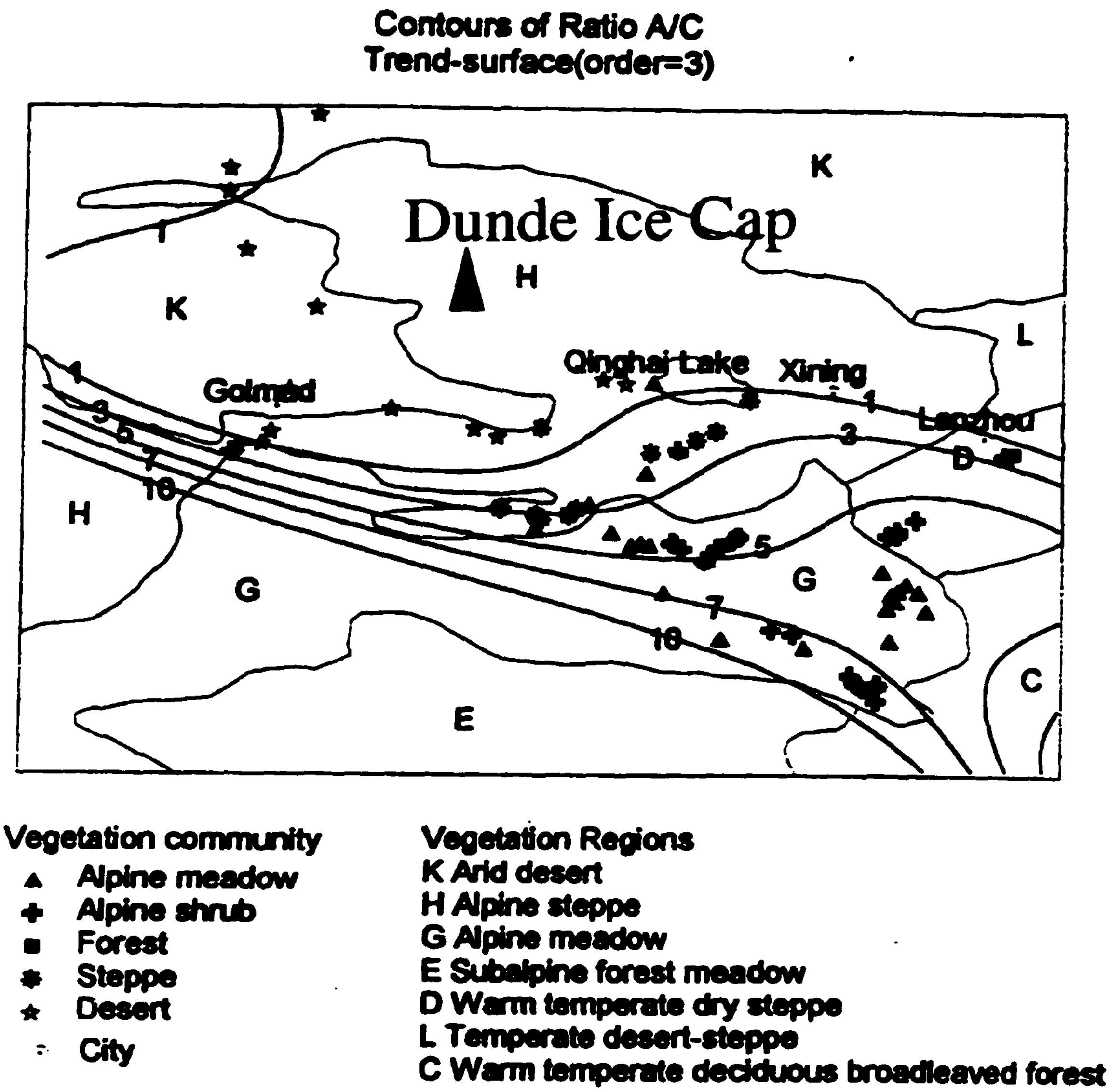


Figure 4.5 Trend surface map of the A/C ratio in the northeastern Qinghai-Tibetan Plateau, based on the 87 surface samples. The interval of each contour is 2.

region. In steppe region, the A/C ratio decreases to between 1 and 5. In desert region, it is less than 1. The A/C ratio decreases from southeast to northwest, corresponding to the trend of decreasing precipitation on the Qinghai-Tibetan as a function of decreasing summer monsoon strength. Therefore, the A/C ratio can be used effectively as a moisture indicator.

4.5 Conclusions

It can be concluded from this surface sample study that major vegetation types of the northern Qinghai-Tibetan Plateau are characterized by distinctive palynological signatures. The results imply that most of pollen come from the local vegetation. The A/C ratio decreases with decreasing precipitation, suggesting that this parameter can be used as a moisture indicator. An increase in A/C ratio in the Dunde ice core may be interpreted as an increase in the strength of the summer monsoon, because most of the precipitation falling in the Dunde region is monsoonal in origin. This knowledge about the relationship between pollen and vegetation is useful in the interpretation of the fossil pollen data from the Dunde Ice Cap.

CHAPTER V

THE RESPONSE OF THE DUNDE ICE-CORE POLLEN RECORDS TO CLIMATE: THE LAST 30 YEARS

5.1 Introduction

The basic idea of using ice-core data to reconstruct a paleoclimatic time series involves the derivation of climatic parameters from proxy data, such as temperature from the oxygen isotope data (Dansgaard et al., 1969; Yao et al., 1997). Knowledge of the relationship between these indicators and climate parameters is usually established on the basis of instrumental observations (Morgan and van Ommen, 1997; van Ommen and Morgan, 1997; Zhang and Yao, 1995a, 1995b; Yao et al., 1997). Therefore, understanding the pollen deposition on ice caps in relation to instrumental climatic observations is essential for interpreting the pollen changes detected in deep ice cores.

High-temporal-resolution pollen analyses of recent ice core samples, especially from non-polar areas, are uncommon, but the few studies available so far have suggested the potential value of such analyses in elucidating the sensitivity of ice-core records to climatic changes (Bourgeois et al., 1985; K-b Liu et al., 1998). High accumulation rates in the upper part of ice cores permit pollen analysis to be conducted at annual or seasonal resolutions to establish the relationship between pollen and climate. In this study, pollen records from a 30-year period (A.D. 1957-1986) are examined to reveal the response of pollen in the Dunde ice core to recent climatic changes. This is the first study on the quantitative relationship between pollen in ice cores and climatic parameters derived from the instrumental observations.

5.2 The Study Region

The Dunde Ice Cap is located in the Qilian mountains on the northern margin of the Qinghai-Tibetan Plateau (Figure 5.1). The ice cap has an area of 60 km^2 and a summit elevation of 5325 m. Steppes and alpine meadows, dominated by *Artemisia* and Cyperaceae, respectively, occur around the ice cap and also to the east and southeast (K-b Liu et al., 1998). Trees are absent in the Dunde region. The mean annual temperature at the Dunde Ice Cap is -7.3°C. Annual accumulation measured from the annual ice layers at the top of the Dunde ice core is approximately 400 mm (water equivalent), but precipitation in the surrounding regions is much lower, about 100-200 mm per year. Over 80% of the snow at Dunde falls in the summer wet season (Thompson et al., 1989).

5.3 Materials and Methods

Ice cores retrieved from the Dunde Ice Cap have already produced a wealth of high-temporal-resolution paleoclimatic

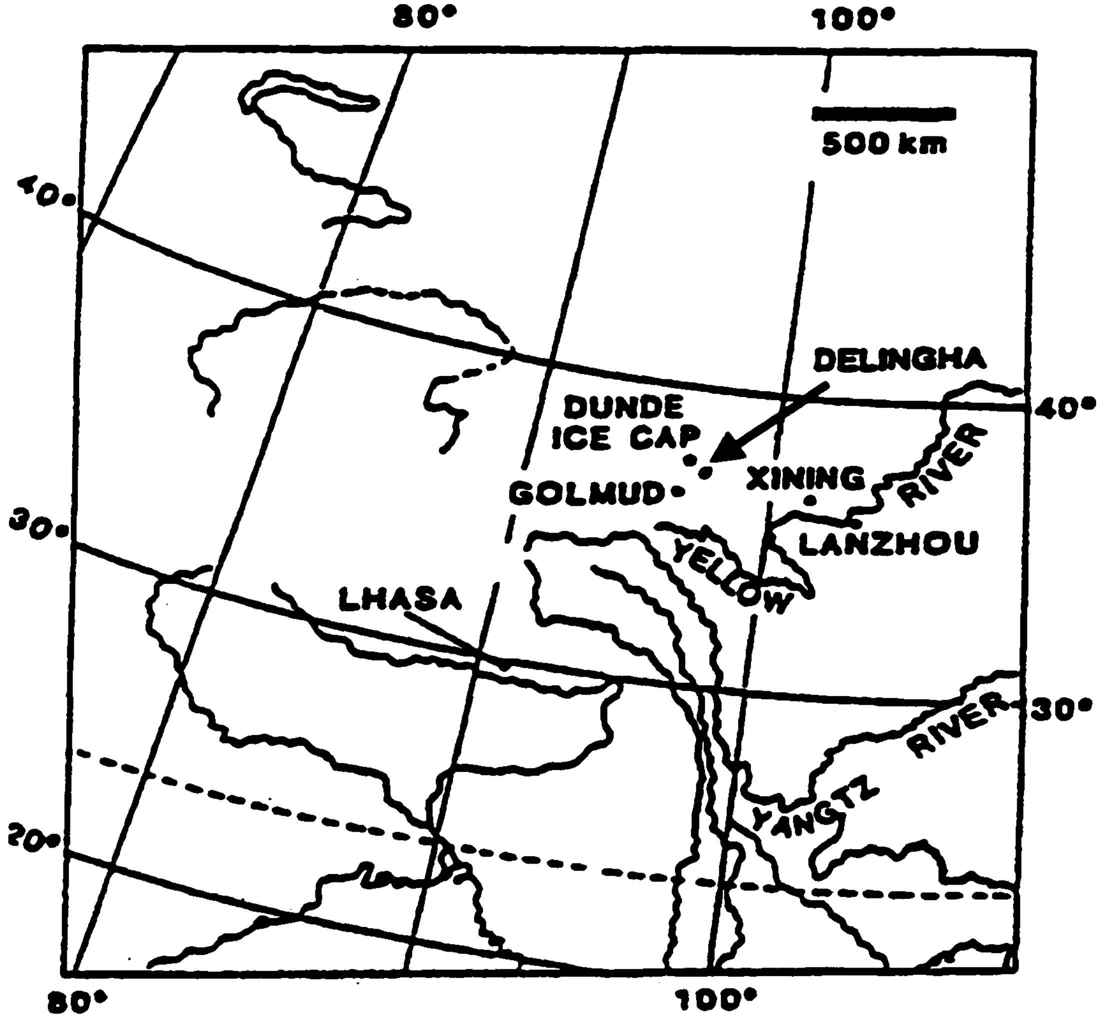


Figure 5.1 Location of the Dunde Ice Cap in relation to the Delingha climatic station (arrow).

data for the northern Qinghai-Tibetan Plateau (Thompson et al., 1989; Yao and Thompson, 1992). Because of the relatively large volume of meltwater needed for the pollen analysis and the depletion of samples for other analytical purposes, samples for the present study have to be derived from different ice cores. Thirty (30) pollen samples (A.D. 1957 to 1986) were derived from two different ice cores: deep core D-3, and shallow core 2. Core D-3 was extracted from the summit of the Dunde Ice Cap, while shallow core 2 was taken near the margin of the ice cap. I used the previously published age model for the Dunde ice cores to provide chronological control for pollen samples (Thompson et al., 1989). The pollen samples for the time interval from 1977 to 1986 come from shallow core 2. For the interval from 1957 to 1976, the pollen samples are from core D-3. Pollen samples ranged in volume from 530 to 850 ml (mean = 650 ml). Although the pollen samples come from different cores, there is no abrupt change of pollen percentages and total pollen concentration across the boundary between these two cores (Figure 5.2). The most remarkable change in the 30-year pollen record is the distinct increase in total pollen concentration values after 1965, accompanied by a notable increase in the percentages of meadow component (Cyperaceae, *Polygonum*) and tree pollen (*Quercus*, Cupressaceae), and a

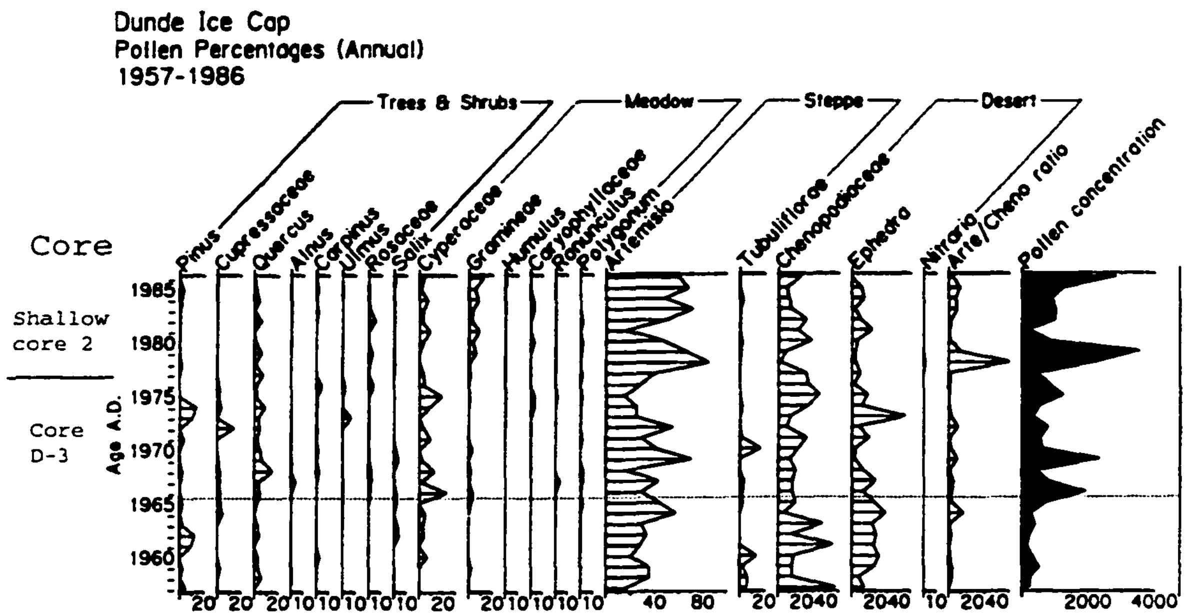


Figure 5.2 Pollen percentage diagram for 30 annual ice layers (1957-1986) in the Dunde Ice Cap. The pollen concentration curve is also presented (far right).

decrease in *Ephedra*. These distinct pollen changes occur within the same core.

Based on modern plant communities, the pollen taxa can be aggregated into four ecological groups: tree and shrub, steppe, desert, and meadow. The tree and shrub group includes *Pinus*, *Picea*, *Quercus*, Cupressaceae, *Betula*, *Alnus*, *Carpinus*, *Ulmus*, *Salix*, and Rosaceae. The steppe group includes *Artemisia*, Tubuliflorae, and *Thalictrum*. The desert group includes Chenopodiaceae, *Ephedra*, *Tamarix*, and *Nitraria*. The meadow group includes Cyperaceae, Gramineae, *Humulus*, Caryophyllaceae, Elaeagnaceae, and *Polygonum*.

5.4 Climatic Stations and Parameters

Meteorological stations are sparse for most parts of the Qinghai-Tibetan Plateau, and most of these stations were established after 1954 (Tao et al., 1991; Riches et al., 1992). Because there is no long-term climatic station in the Dunde area, it is necessary to use data from climatic stations within a broader geographical region. X.D. Liu et al. (1998) used 48 climatic stations, and most of them located on the eastern part of Qinghai-Tibetan Plateau, to study the spatial correlation among these stations. They found that the annual temperature records showed the same trend on the Qinghai-Tibetan Plateau. Figure 5.3 shows distribution of the correlation coefficients between annual mean surface air-temperature of the Qumarle station and

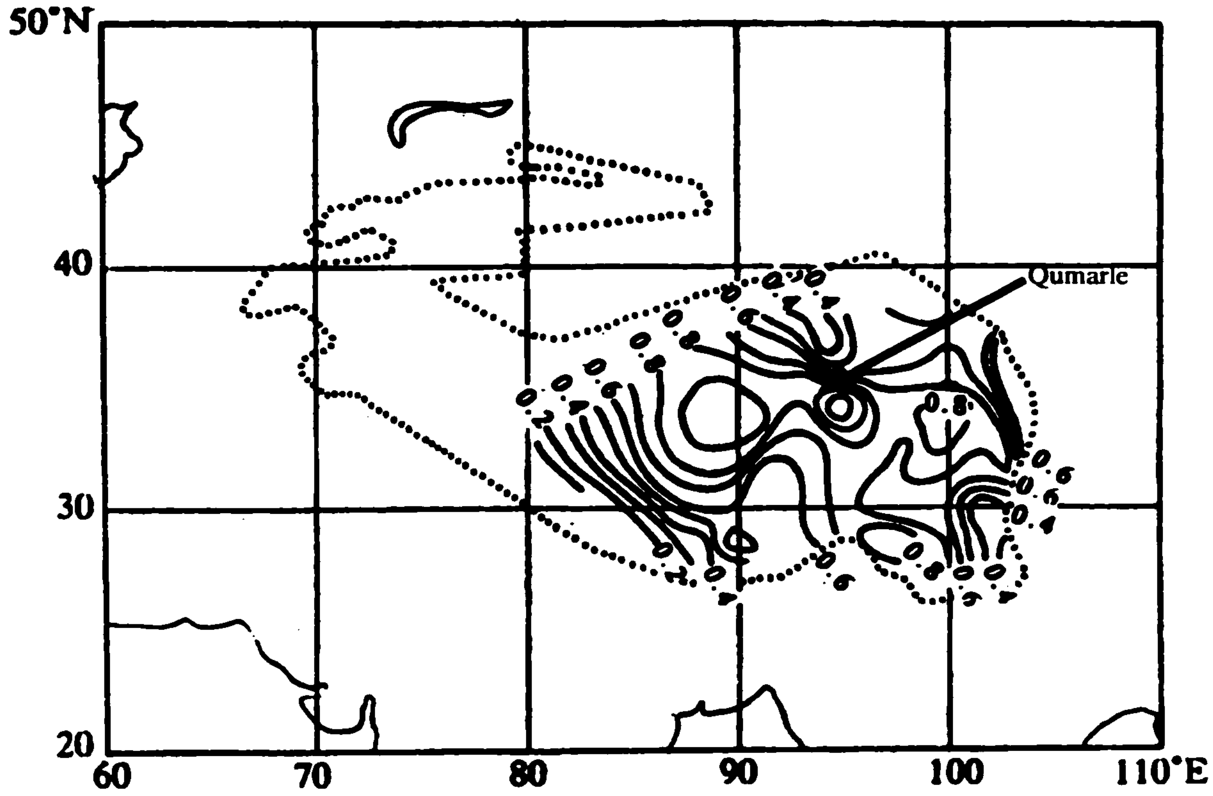


Figure 5.3 Distribution of correlation coefficients between annual mean surface air temperature of station Qumarle and 47 stations on the Qinghai-Tibetan Plateau (After X.D. Liu et al., 1998). The dotted line is the 4000 m contour line of the Qinghai-Tibetan Plateau. The arrow points to Qumarle.

those of 47 stations on the Qinghai-Tibetan Plateau. The result shows that the closer between two stations the higher the correlation coefficient, and vice versa. Therefore, the climatic station at Delingha, which is the closest to the Dunde (Figure 5.1), was chosen for establishing the relationship between pollen and climate.

The climatic variables used to study the relationship between pollen and climate should be ecologically meaningful. Precipitation and temperature are very important factors for plant growth and vegetation distribution. In this study, precipitation (including annual, summer, and spring precipitation) and temperature (including annual, summer, and winter temperature) were used to establish the relationship between pollen and climate.

5.5 Correlation between Pollen and Precipitation

Figure 5.4 shows the relationship between pollen and annual precipitation. There is no significant correlation between percentages of tree pollen and annual precipitation (Table 5.1). It is reasonable because trees are absent in the Dunde area. The tree pollen is developed from long-distance transport. It may reflect wind direction (Bourgeois et al., 1985; Bourgeois, 1990b; Sun et al., 1993, 1994; Nichols et al., 1978).

Statistical results also show that total pollen concentration, percentages of meadow pollen, and percentages

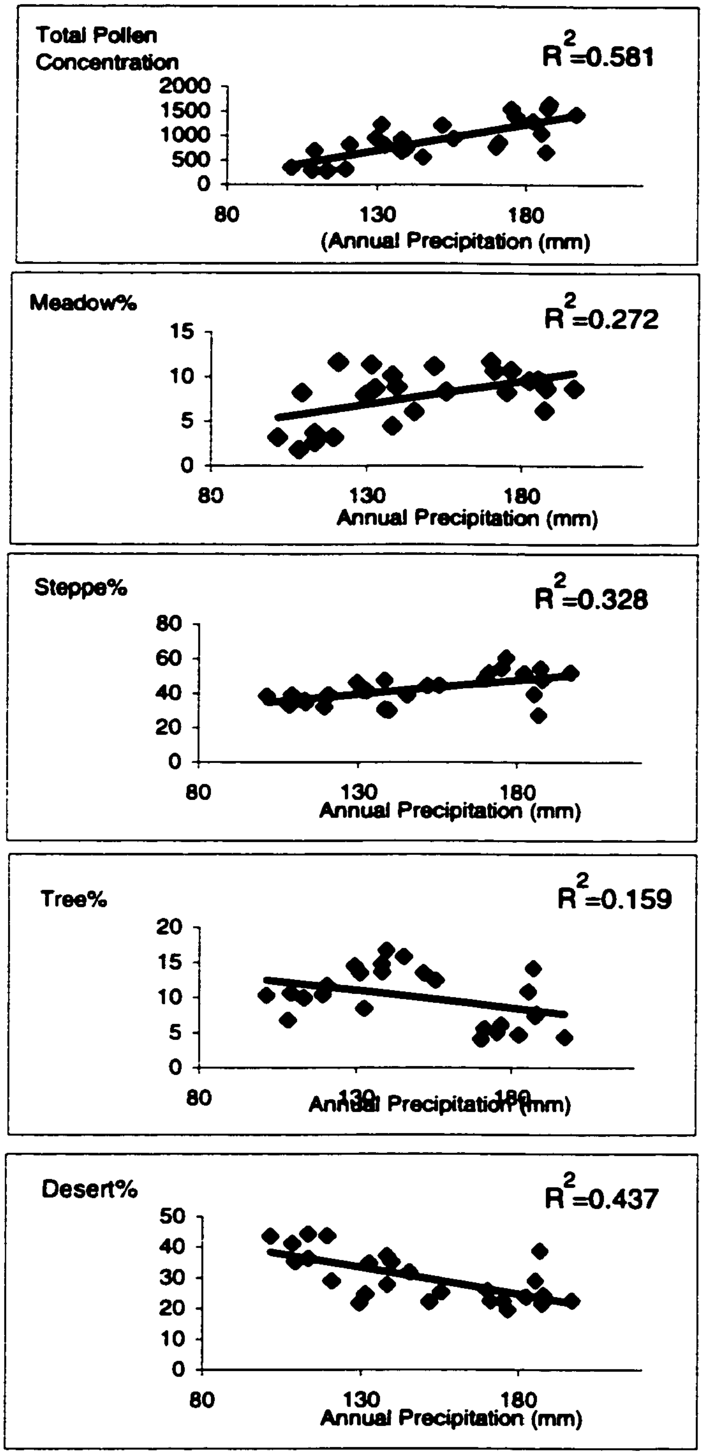


Figure 5.4 The relationship between pollen and annual precipitation. The climate data is from Delingha. The line is regression line.

Table 5.1 Correlation between pollen and annual precipitation for the last 30 years.

Pollen types	R	R-square	F-value	Significance level
Total pollen concentration	0.762	0.581	33.23	0.001*
Meadow	0.522	0.272	8.975	0.006*
Steppe	0.571	0.328	11.609	0.002*
Desert	-0.662	0.439	18.770	0.001*
Tree	-0.398	0.159	3.527	0.054

* significance level (<0.05).

of steppe pollen are positively correlated with annual precipitation, while percentage of desert is negatively correlated with annual precipitation at the 95% significance level (Table 5.1). These results suggest that an increase in annual precipitation will result in an increase in the percentages of meadow and steppe but a decrease in the percentages of desert pollen. Therefore, an increase in percentages of meadow and steppe in the ice core may indicate an increase in precipitation, while an increase in percentages of desert pollen may imply a decrease in annual precipitation.

Figure 5.5 shows the relationship between pollen and summer precipitation. The summer precipitation is a total of precipitation in June, July, and August (JJA). The summer precipitation is an important parameter because it accounts for 80% of the annual total in the Dunde region; it is a good indicator of the summer monsoon (Thompson et al., 1989). The statistical results show that the percentages of steppe and percentages of tree pollen have no significant correlation with summer precipitation (Table 5.2). The total pollen concentration and percentages of meadow are positively correlated with summer precipitation, while percentages of desert are negatively correlated with summer precipitation at a 95% significance level. The results indicate that an increase in summer precipitation will

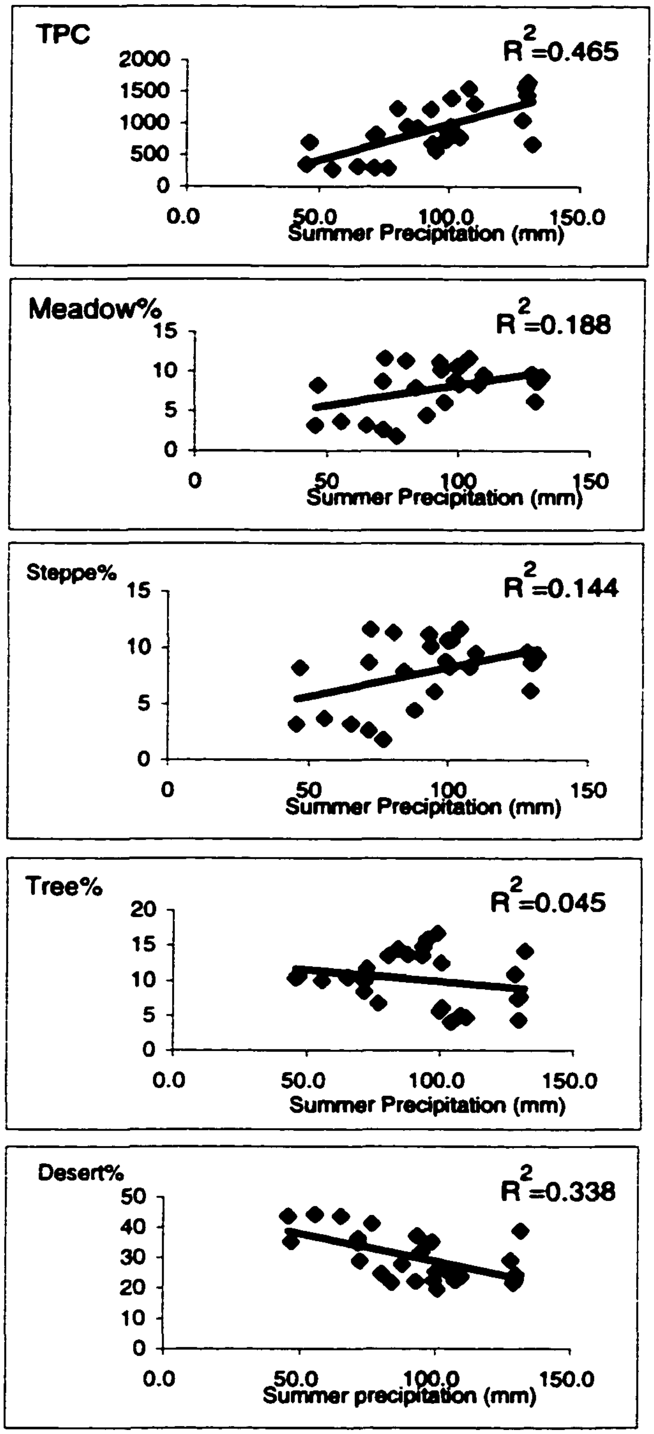


Figure 5.5 The relationship between pollen and summer precipitation. The climate data is from Delingha. The line is regression line.

Table 5.2 Correlation between pollen and summer precipitation for the last 30 years

Pollen types	R	R-square	F-value	Significant level
Total pollen concentration	0.682	0.465	20.871	0.001*
Meadow	0.433	0.188	5.54	0.027*
Steppe	0.379	0.144	4.038	0.056
Desert	-0.581	0.338	12.237	0.002*
Tree	-0.212	0.045	1.128	0.299

* significance level (<0.05).

result in an increase in total pollen concentration and meadow pollen. Therefore, an increase in total pollen concentration and percentages of meadow in the ice core may suggest an increase in precipitation.

Figure 5.6 shows the relationship between pollen and spring precipitation. The spring precipitation is a total of precipitation in March, April, and May (MAM). The statistical results show that percentages of meadow and percentages of steppe are positively correlated with spring precipitation, while the percentages of desert are negatively correlated with spring precipitation at a 95% significance level (Table 5.3). But there is no significant correlation between total pollen concentration and percentages of tree and spring precipitation (Table 5.3).

A climatic response surface is a nonlinear function describing the way in which the abundance of taxa depends on the joint effects of two or more environmental variables (Bartlein et al., 1986). Figure 5.7 shows the response surface of total pollen concentration to summer and spring precipitation. If given a lower spring precipitation, total pollen concentration will increase with summer precipitation slowly, while if given a higher spring precipitation, total pollen concentration will increase greatly. It means that the total pollen concentration is affected by not only summer precipitation but also spring precipitation. The

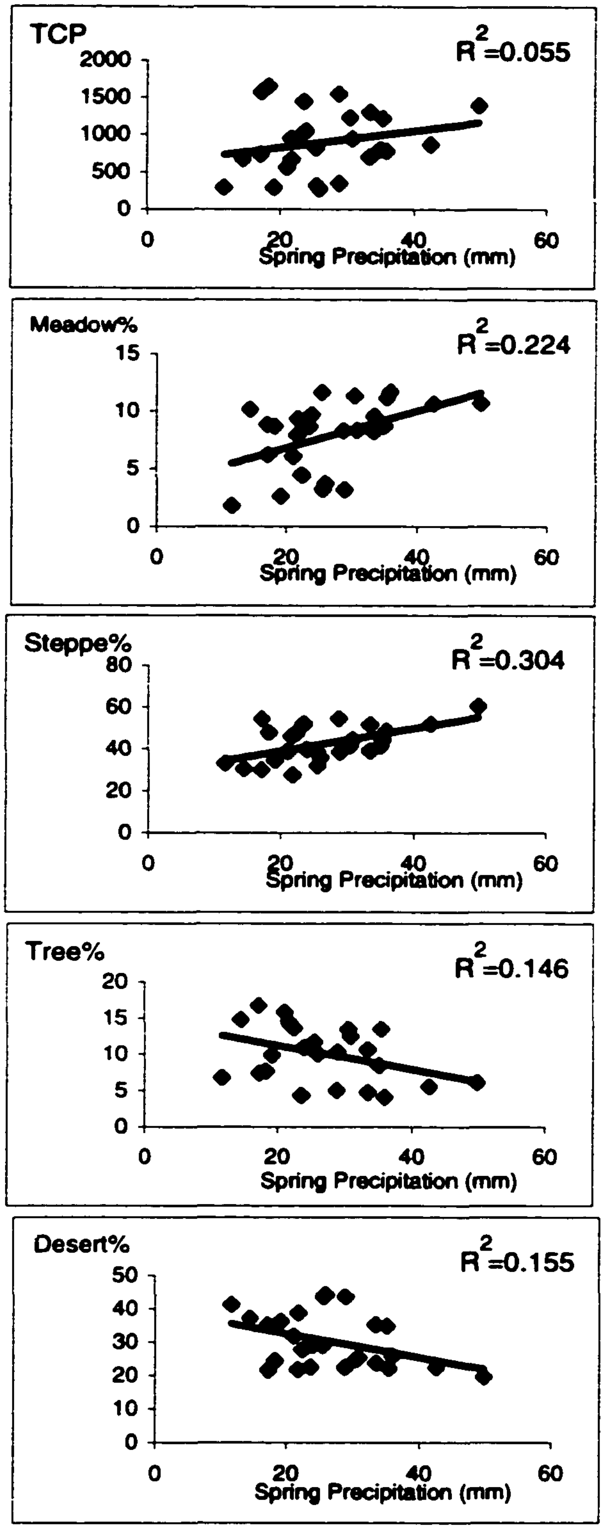


Figure 5.6 The relationship between pollen and spring precipitation. The climate data is from Delingha. The line is regression line.

Table 5.3 Correlation between pollen and spring precipitation for the last 30 years.

Pollen types	R	R-square	F-value	Significant level
Total pollen concentration	0.234	0.055	1.396	0.249
Meadow	0.474	0.224	6.938	0.015
Steppe	0.551	0.304	10.48	0.004
Desert	-0.394	0.155	4.419	0.046
Tree	-0.382	0.146	4.094	0.054

* significance level (<0.05)

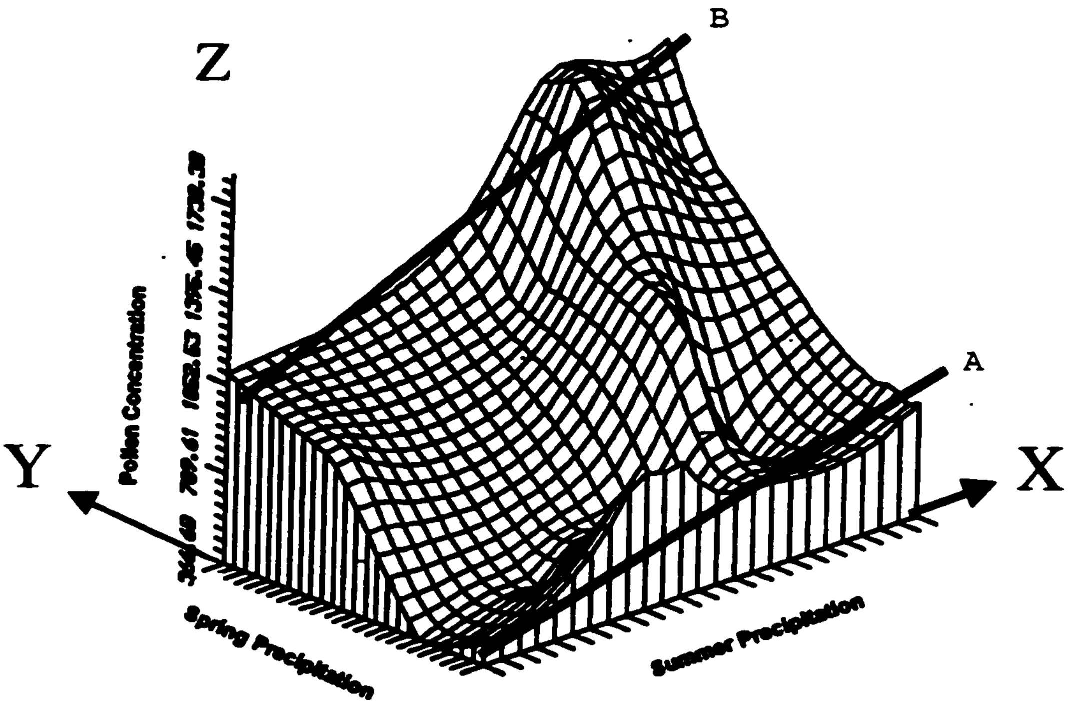


Figure 5.7 Response surface of pollen concentration and climate parameters. X-axis = summer (JJA) precipitation, Y-axis = spring (MAM) precipitation; Z-axis = pollen concentration. Climatic data based on a meteorological station of Delingha. Slope of line A is smaller than that of line B.

results show that summer (JJA) and spring (MAM) precipitation has interactive effects on total pollen concentration, although the spring precipitation has no linear relation with total pollen concentration. Therefore the higher spring precipitation is favorable to plant growth by increasing soil moisture.

In summary, different pollen types shows different responses to precipitation in the Dunde region. Although the relationship between pollen and precipitation is complicated, it can be estimated that an increase in precipitation (e.g. annual, summer, and spring precipitation) is favorable to meadow and steppe growth, resulting in a high total pollen concentration. An increase in the percentages of desert pollen in the ice core may imply a decrease in moisture.

5.6 Correlation between Pollen and Temperature

Figure 5.8 shows the relationship between pollen and summer temperature. The summer temperature is an average of June, July, and August (JJA) temperatures. The statistical results show that there is no significant correlation between percentages of steppe and tree and summer precipitation at a 95% significance level (Table 5.4). However, the total pollen concentration and percentage of meadow are negatively correlated with summer temperature, while percentage of desert is positively correlated with

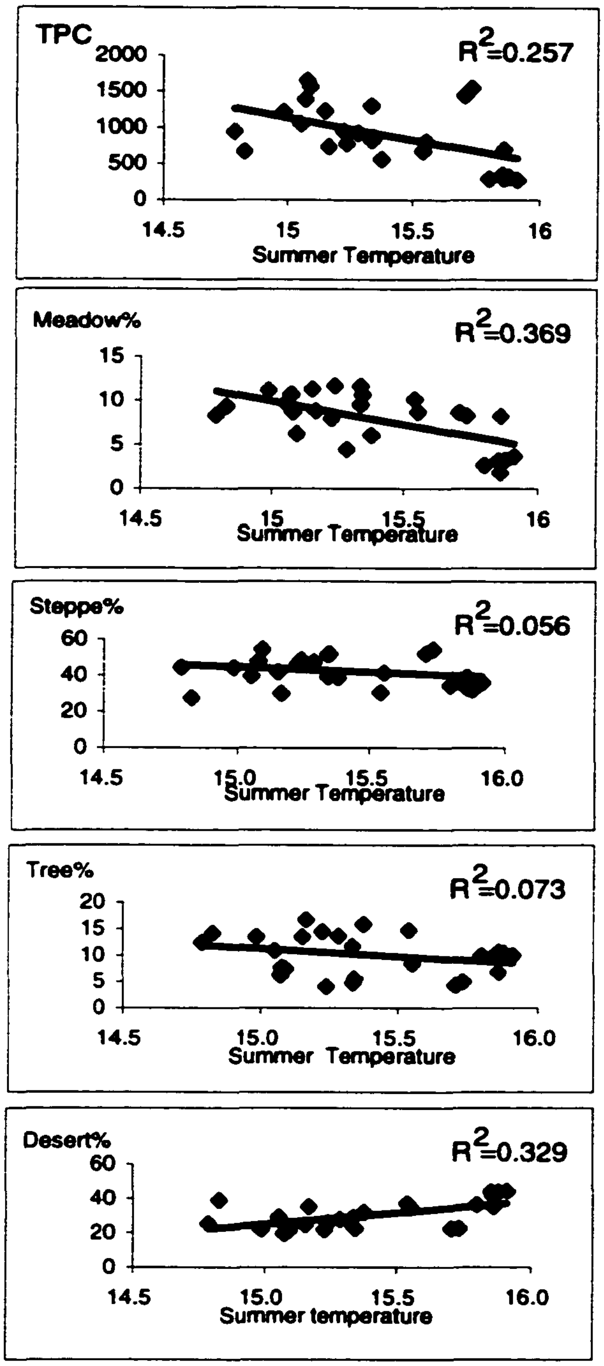


Figure 5.8 The relationship between pollen and summer temperature. The climate data is from Delingha. The line is regression line.

Table 5.4 Correlation between pollen and summer temperature for the last 30 years.

Pollen types	R	R-square	F-value	Significance level
Total pollen concentration	-0.507	0.257	8.302	0.008*
Meadow	-0.608	0.369	14.046	0.001*
Steppe	-0.236	0.056	1.414	0.246
Desert	0.574	0.329	11.79	0.002*
Tree	-0.269	0.073	1.877	0.183

*significance level (<0.05)

summer temperature at the 95% significance level. It is reasonable because an increase in summer temperature will decrease the effective moisture during growing season, and a decrease in summer temperature will reduce evapotranspiration and increase the effective moisture; these conditions are favorable to plant growth.

Figure 5.9 shows the relationship between pollen and winter temperature. The winter temperature is an average of December, January, and February (DJF) temperatures. The statistical results show that there is no significant correlation between the percentages of trees and winter temperature at the 95% significance level (Table 5.5). However, the total pollen concentration and percentages of meadow and steppe are positively correlated with winter temperature at the 95% significance level. Warm winters may melt snow earlier and increase soil moisture in the spring to favor to meadow and steppe growth, resulting in high total pollen concentration.

Figure 5.10 shows response of total pollen concentration to summer temperature and winter temperature. Total pollen concentration is strongly correlated with summer temperature and weakly with winter temperature. It suggests that summer temperature is one of the main factors

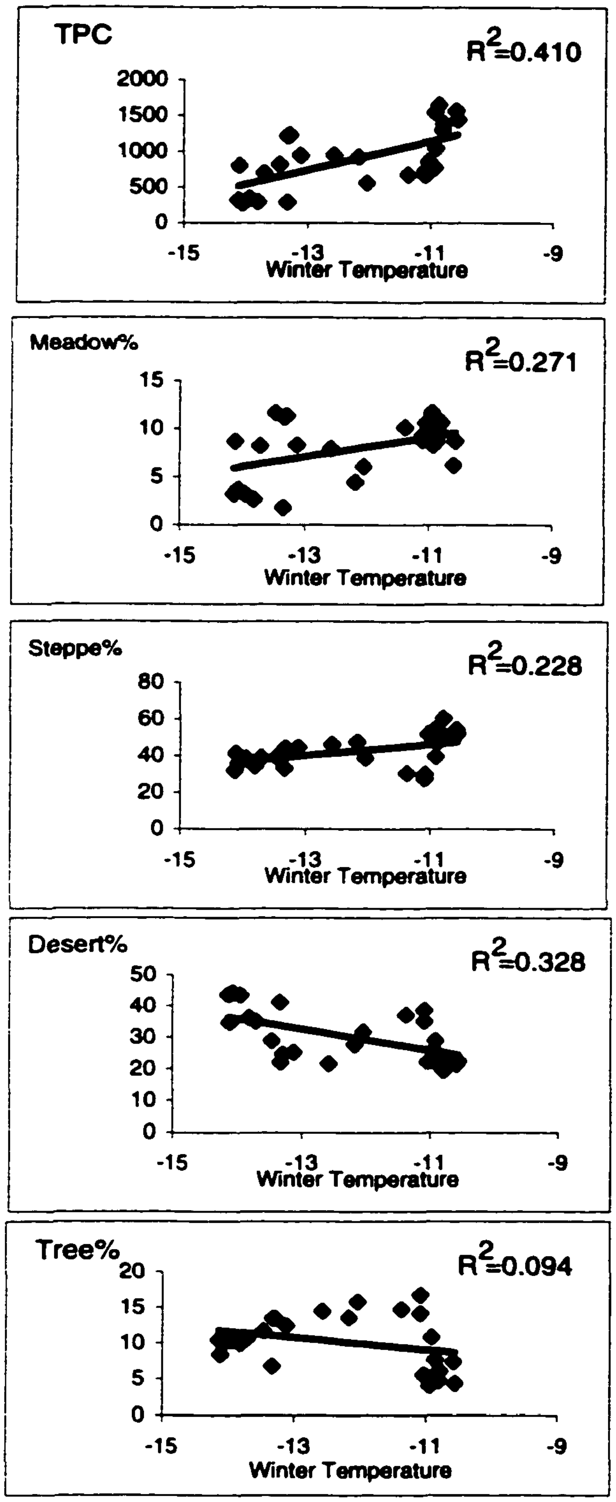


Figure 5.9 The relationship between pollen and winter temperature. Climate data is from Delingha. The line is regression line.

Table 5.5 Correlation between pollen and winter temperature for the last 30 years.

Pollen types	R	R-square	F-value	Significance level
Total pollen concentration	0.64	0.410	16.668	0.001
Meadow	0.460	0.271	6.426	0.018
Steppe	0.478	0.228	7.092	0.014
Desert	-0.573	0.328	11.709	0.002
Tree	-0.306	0.094	2.48	0.128

* significance level (<0.05)

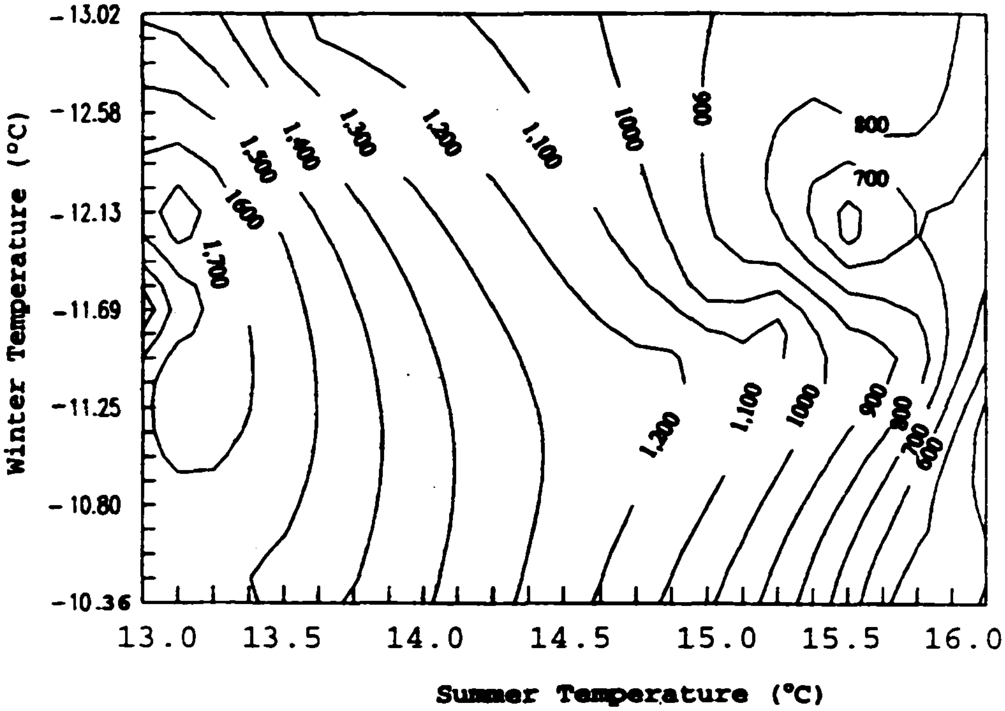


Figure 5.10 Contours of pollen concentrations plotted against summer and winter temperatures. X-axis = mean summer temperature (JJA); Y-axis = winter temperature (DJF). All climatic data are from the meteorological station at Delingha.

affecting total pollen concentration. Bourgeois (1990b) also proposed that summer temperature was an important factor affecting total pollen concentration in the polar region.

Figure 5.11 shows the relationship between pollen and annual temperature. The statistical results show that there is no significant correlation between percentages of meadow and tree and annual temperature at the 95% significance level (Table 5.6). The percentages of desert pollen are positively correlated with annual temperature. It suggests that high annual temperature will reduce soil moisture, thus favoring to xerophilous plants.

In summary, temperature influences both pollen percentages and total pollen concentration by affecting soil moisture. An increase in summer temperature will reduce effective moisture, favoring xerophilous plants. A decrease in summer temperature will increase effective moisture, favoring moisture-loving plants such as those dominanted in the meadows.

5.7 Conclusions

This is the first quantitative study of the relationship between ice-core pollen and climatic parameters. Total pollen concentration, percentages of meadow pollen, and percentages of steppe pollen are positively correlated with annual precipitation, summer precipitation, and winter temperature, while negatively

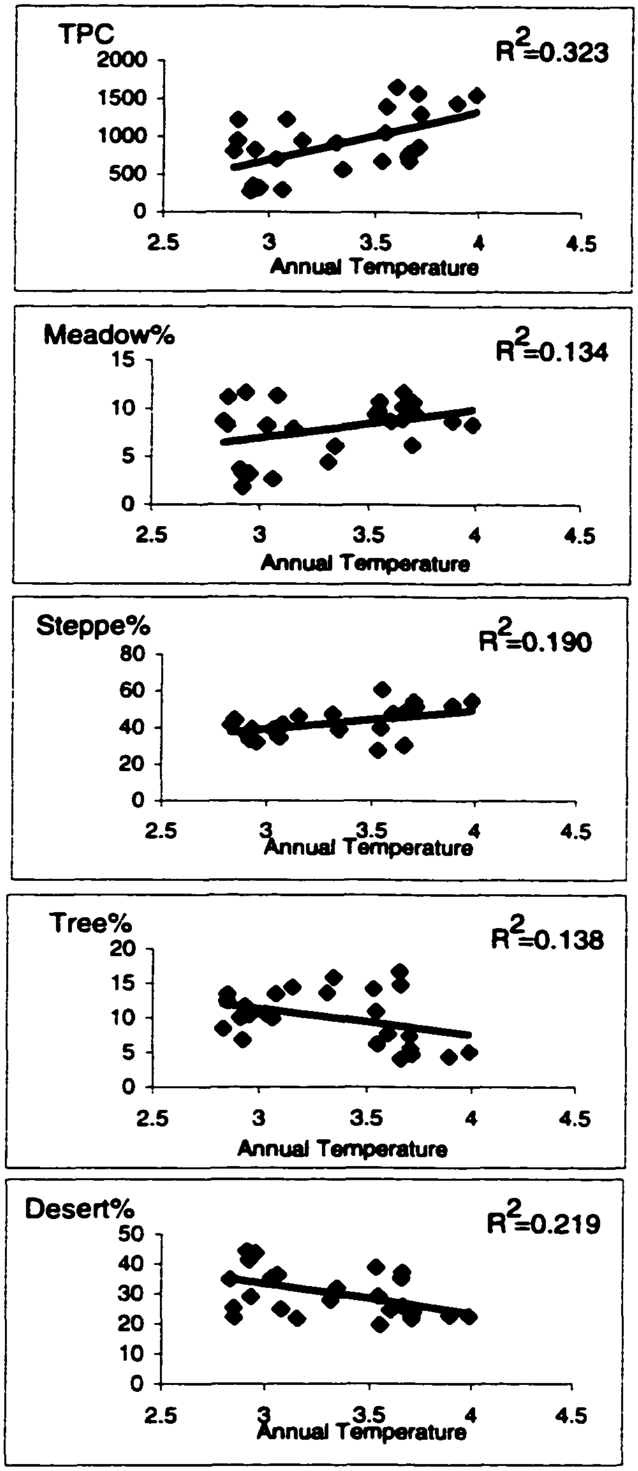


Figure 5.11 The relationship between pollen and annual temperature. The climate data is from Delingha. The line is regression line.

Table 5.6 Correlation between pollen and annual temperature for the last 30 years.

Pollen types	R	R-square	F-value	Significance level
Total pollen concentration	0.568	0.323	11.43	0.002
Meadow	-0.366	0.134	3.707	0.066
Steppe	-0.436	0.190	5.626	0.026
Desert	0.468	0.219	6.727	0.016
Tree	-0.372	0.138	3.851	0.061

* significance level (<0.05)

correlated with summer temperature. The percentages of tree and desert pollen are positively correlated with annual temperature and summer temperature, while negatively correlated with annual precipitation and summer precipitation. Therefore, an increase in total pollen concentration and in the percentages of meadow and steppe pollen indicate an increase in effective moisture, while an increase in desert pollen may indicate dry condition. The correlation between pollen and climatic parameters can be used to interpret pollen record of deep ice cores.

CHAPTER VI

COMPARISON OF POLLEN RECORDS BETWEEN PARALLEL ICE CORES

6.1 Introduction

Since the first ice-core study was published in 1969 (Dansgaard, 1969), a lot of physical and geochemical data have been used to provide a variety of unique climatic information. Because more and more physical and geochemical measurements are used to extract the climatic information, parallel cores have to be drilled. For example, three deep cores were extracted from the summit of the Dunde Ice Cap. However, there are few published studies that compare the stratigraphies between parallel cores.

Pollen studies of lake-sediments suggest that sediment focusing can cause horizontal variations in pollen concentration and pollen influx in different parts of a lake (Davis and Ford, 1982). In conical basins, which are more typical of real lakes, sediment is generally focused toward the center. As a lake basin fills in, younger sediment may be spread out over a larger area of the lake bottom, thus accumulating in thinner layers. Under these circumstances the sediment core can register a misleading decline in influx, even though inputs to a lake as a whole remains constant. Therefore, pollen concentration and influx show inconsistent values in different parts of a lake.

However, the depositional environments and processes on an ice cap are different from those in a lake. Because of ice flow and ice-melt in the margin of an ice cap, the annual ice layers are thicker in the summit than on the margin. Does it influence pollen records from different ice cores in the Dunde Ice Cap?

6.2 Data Source

The annual pollen records of two ice cores from the summit of the Dunde Ice Cap (core D-1 and core D-3) (Figure 3.6) spanning from A.D. 1720 to 1936 (266-50 BP) were compared with each other. Overlapping samples are available only from these two cores, because samples from other parts of these and other cores have been used up for other analyses.

6.3 Methods

Two statistical methods are used in this study. First, a T-test (using SAS 8.0) was used to test whether sampling sites have no effect on the pollen records from two sites (null hypothesis). Second, cross-correlation was used to test the correlation between two cores. This method has been used in tree-ring study for dating or remote cross-dating (Wu, 1990), palynological studies (Green, 1981, 1983), and forest fire studies (Chen, 1990). The formulas have been introduced by Green (1981).

6.4 Results

Artemisia and Chenopodiaceae together account for about 80% of the pollen sum in most of the pollen samples from the Dunde ice cores. In this study, four variables are used: *Artemisia* %, *Ephedra* %, and Chenopodiaceae %, and total pollen concentration.

Figure 6.1 presents the downcore variations of total pollen concentrations during the period 266-50 BP from core D-1 and core D-3. The mean of total pollen concentration (1305.4±125) at core D-1 is not significantly different from the mean of total pollen concentration (1015.9±93) at core D-3 at the 95% significance level (Table 6.1). The statistical result may imply that pollen input at the two cores could be similar. However, the correlation coefficient between the two cores using sample-by-sample comparison is low (r=0.03), and there is no significant correlation between the two cores (p=0.975).

Figure 6.2 presents the *Artemisia* variations. The mean of percentages of *Artemisia* pollen (39.83±2.28) at core D-1 is not significantly different from the mean of percentages *Artemisia* pollen (46.82±2.38) at core D-3 at 95% significance level (Table 6.1). But the correlation coefficient (r) on a sample-by-sample basis is less than 0.1 (-0.025). However, using five-sample running mean,

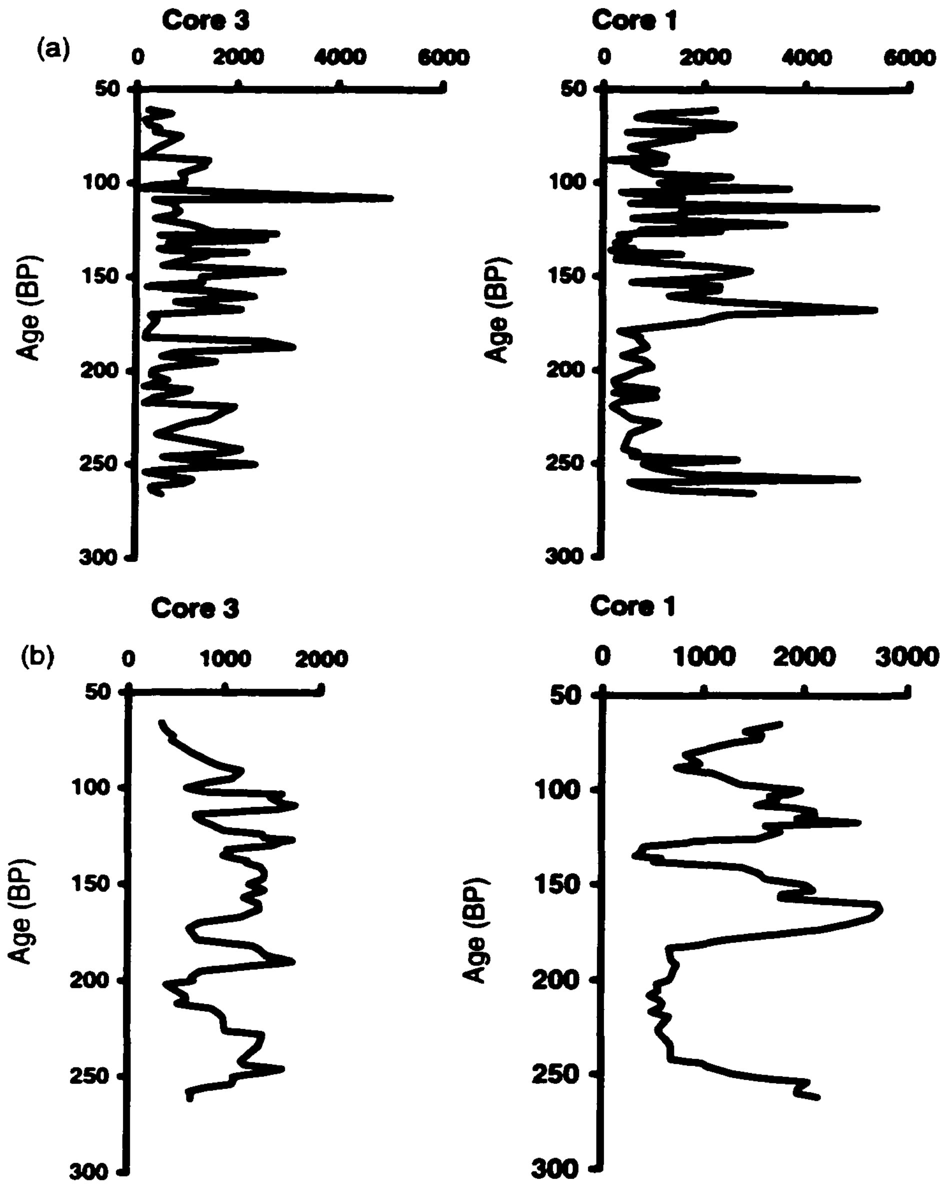


Figure 6.1 Comparison of total pollen concentration curves between core D-3 and core D-1. (a) upper two curves are sample-by-sample plots; (b) lower two curves are five-year running means.

Table 6.1 Paired sample T-test results.

Variables	T-value	Significance level (two tailed)
Artemisia	1.023	0.342
Ephedra	1.115	0.268
Chenopodiaceae	1.425	0.158
Total pollen concentration	1.850	0.068

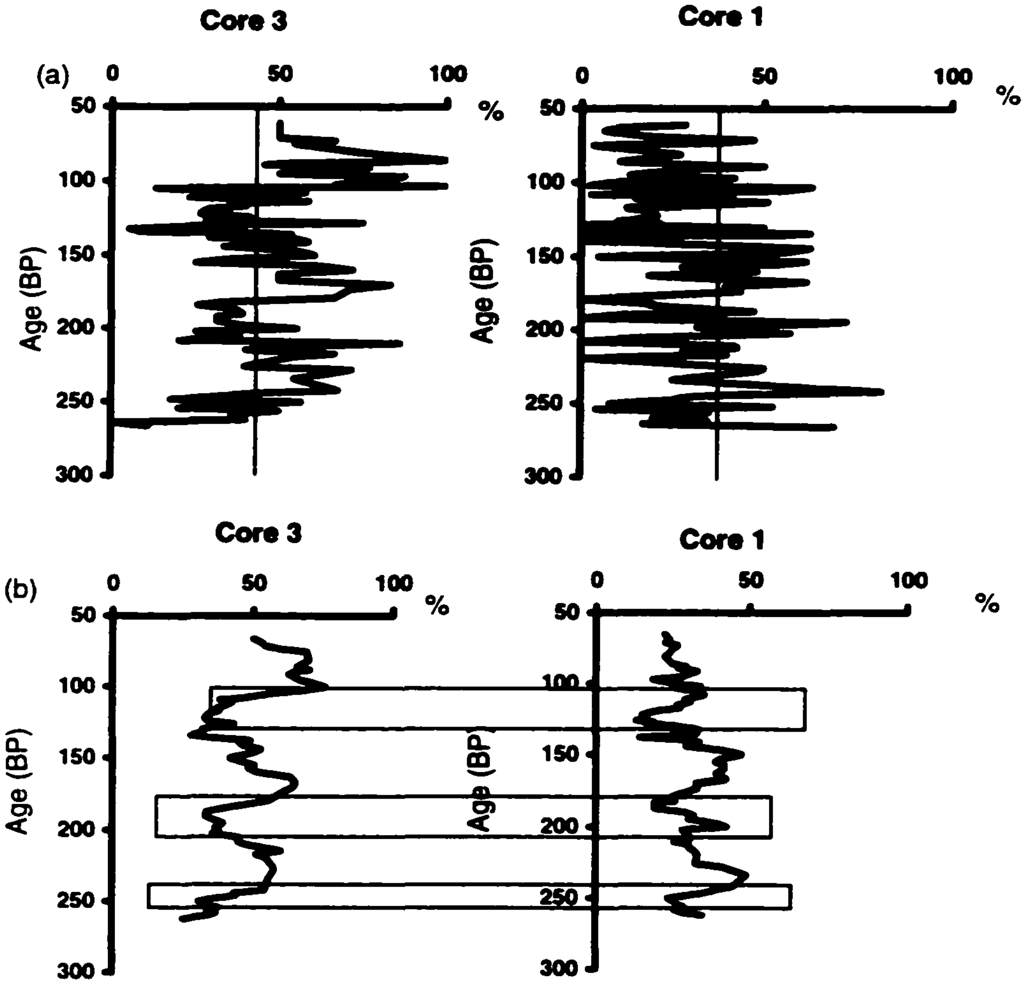


Figure 6.2 Comparison of *Artemisia* pollen percentages between core D-3 and core D-1. (a) upper two curves are sample-by-sample. (b) lower two curves are five-year running means.

Figure 6.2b shows that three lower percentages of *Artemisia* pollen in core D-1 are correlated with three lower percentages of *Artemisia* pollen in core D-3, probably implying the similar trends in the two ice cores. The correlation coefficient (r) also increases to 0.323, although it is still low.

Figure 6.3 shows the Chenopodiaceae variations. The means of percentages of Chenopodiaceae pollen from the two core are 20.94±1.7 and 18.05±1.6, respectively. There is not significant difference between the two cores at the 95% significance level (Table 6.1). But the two cores show a negative correlation (r=-1.96) sample-by-sample comparison (Figure 6.3).

Figure 6.4 shows variations of *Ephedra* pollen. There is no significant difference between the two cores at the 95% significance level (Table 6.1). The means of percentages of *Ephedra* pollen are 12.66±1.56 and 10.26±1.19, respectively. The *Ephedra* pollen shows a negative correlation based on sample-by-sample comparison. However, using the five-year running mean, the correlation coefficient changed from negative (-0.126) to positive (0.386), probably indicating that the *Ephedra* has similar trends at the lower parts of the two ice cores.

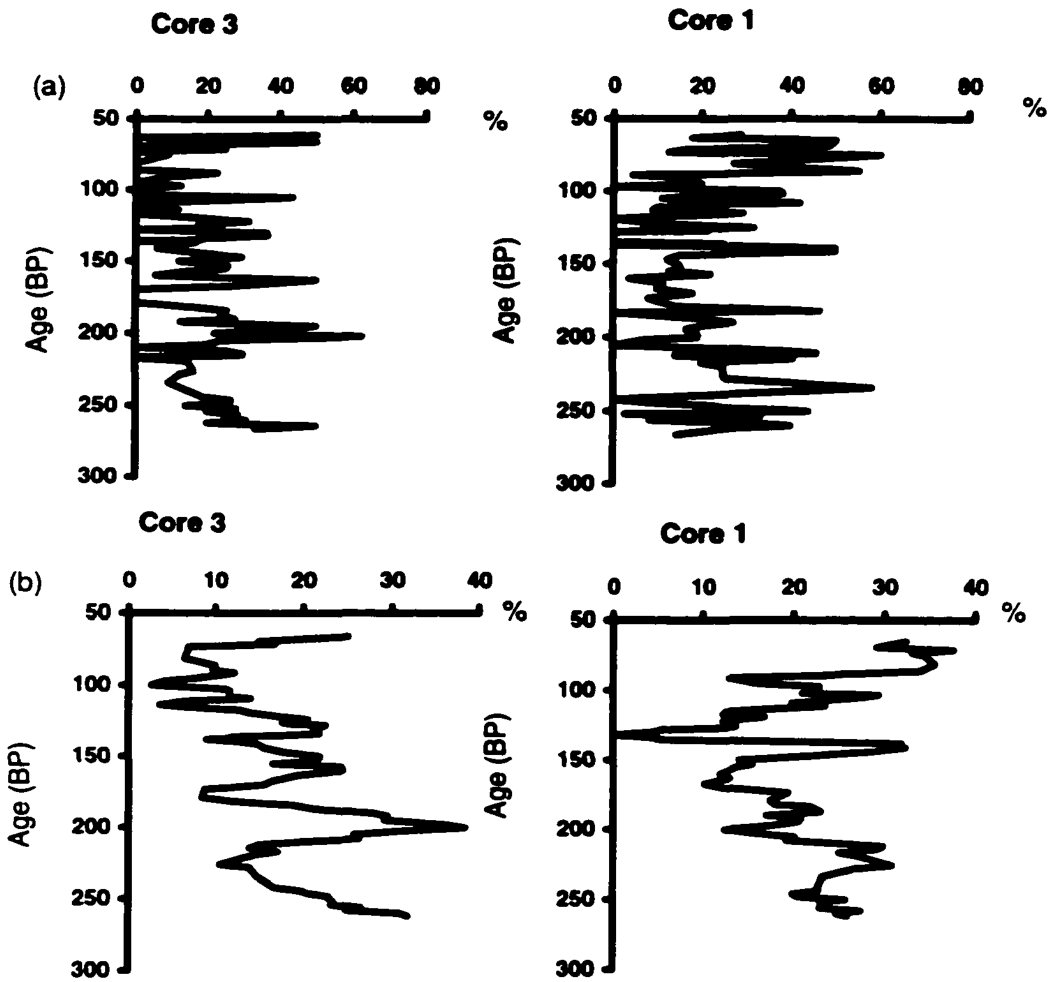


Figure 6.3 Comparison of Chenopodiaceae pollen percentages between core D-3 and core D-1. (a) upper two curves are sample-by-sample plots; (b) lower two curves are five-year running means.

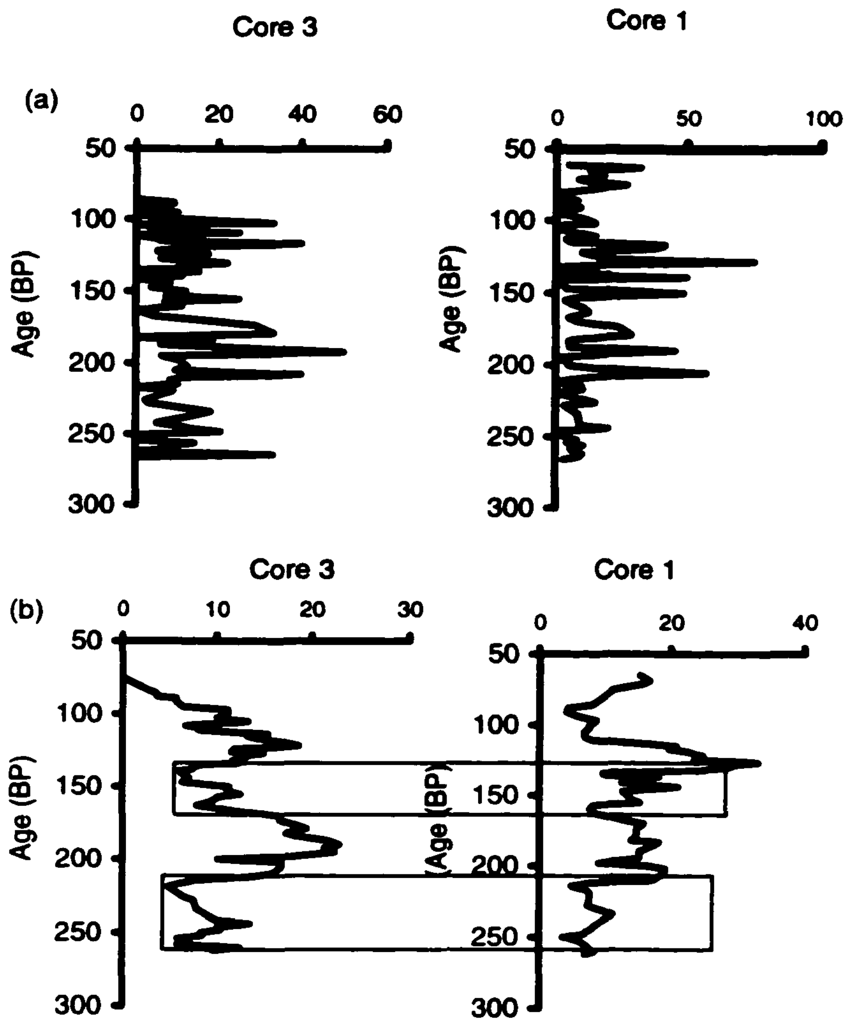


Figure 6.4 Comparison of *Ephedra* pollen percentages between core D-3 and core D-1. (a) upper two curves are sample-by-sample plots; (b) lower two curves are five-year running means.

6.5 Discussions

The pollen data suggest no significant difference of means between the two cores for these variables (Table 6.1). This probably indicates that variations of main pollen types are similar between cores in Dunde. But, correlation coefficients are low for all these variables between the two ice cores based on sample-by-sample comparison.

Cross-correlation analysis is a good method to test whether there are sample shifts between two ice cores. The results of cross-correlation analysis (Figure 6.5) show that Chenopodiaceae, *Artemisia*, and total pollen concentration have negative correlation at zero (no sample shift). The cross-correlation coefficient increases and becomes positive when the samples in core D-3 are lag or ahead in core D-1. It probably indicates that the two cores have sample shifts. One possible explanation is missing some layers. The annual layer for any given year depends upon the specific parameter used for layer identification (Yao and Thompson, 1992). Missing concentration peaks for any given parameters in some years can contribute a significant uncertainty to the determination of change of pollen in the ice core. This is similar to the effect of missing tree-rings on cross-correlating dendrochronologies. Ice-core studies from the polar area also shows the shifts between the GISP2 and GRIP Greenland ice cores (The GRIP core was drilled 28 km east of

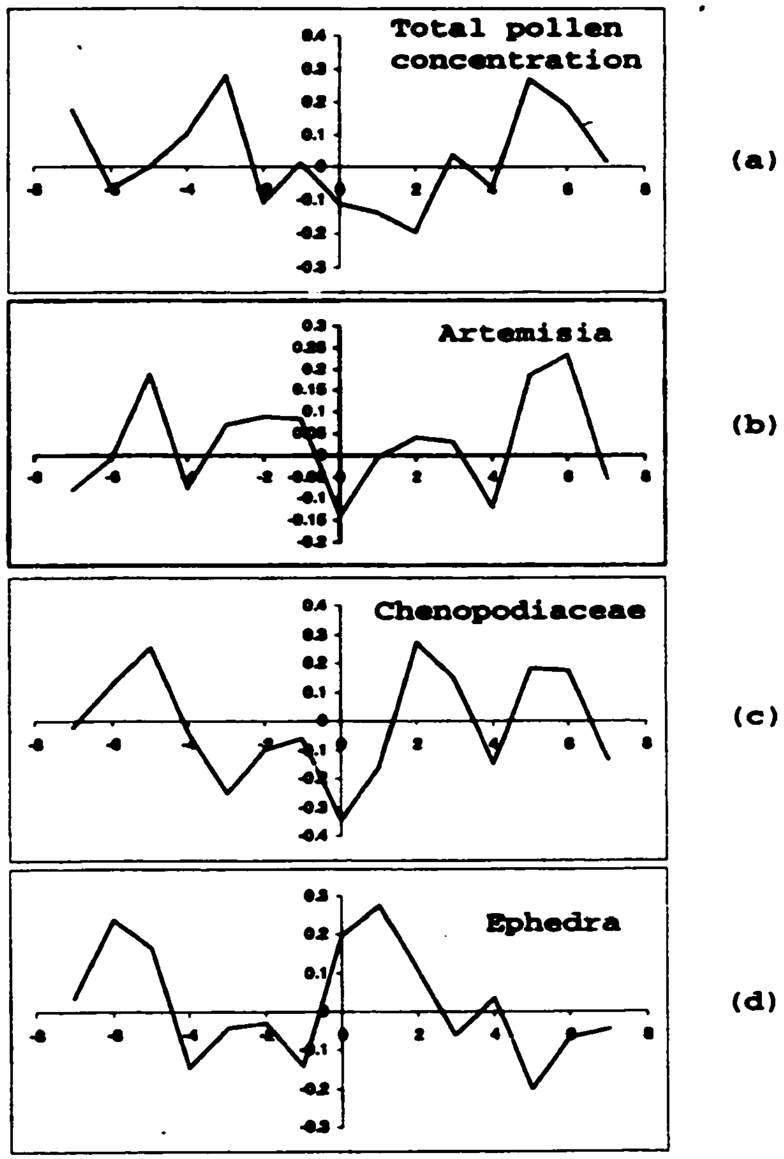


Figure 6.5 Cross-correlation analysis from core D-1 and core D-3. (a) total pollen concentration; (b) *Artemisia*; (cc) Chenopodiaceae; (d) *Ephedra*.

the GISP2) (Grootes et al., 1993), although oxygen isotope records between the two cores showed similar trends throughout the last glacial period.

Another possibility is that the different sampling methods may cause a difference between parallel ice cores. Thompson et al. (1993) presented 50-yr averages of $\delta^{18}O$ for the last 12,000 yr for two parallel cores, core D-1 and core D-3. Both cores were drilled at the summit on the Dunde Ice Cap. Although the general trends of the two $\delta^{18}O$ records are comparable, there are also statistically significant differences between the two cores. Thompson et al. (1993) proposed that the sampling method might cause the difference. Core D-1 was returned to OSU (Ohio State University) entirely as bottled samples, while core D-3 were returned as ice. Thus, core D-3 was sectioned into much smaller samples which provided a higher time resolution. In core D-3 the most recent 50-yr average represents 384 samples and the lower 50-yr averages represent 4 to 10 samples. By contrast, in core D-1 the most recent 50-yr average includes 320 samples and the lower 50-yr averages are based on 2 to 5 samples.

6.6 Conclusions

There is no significant difference between the two ice cores for four variables: *Artemisia* %, Chenopodiaceae %, *Ephedra* %, and total pollen concentration. This suggests that the pollen records have similar trends on the Dunde ice cores. The cross-correlation analysis shows that the percentages of *Artemisia*, *Ephedra*, Chenopodiaceae, and total pollen concentration are negative correlation at zero (no sample shift). There are two possible reasons that may cause the negative correlation: missing layers or different sampling methods. The correlation coefficient could be increased by shifting the samples in one core.

CHAPTER VII

CLIMATIC CHANGES DURING THE LAST 2000 YEARS: POLLEN STUDY OF THE DUNDE ICE CORE

7.1 Introduction

Knowledge about the climatic history of the Qinghai-Tibetan Plateau is important for understanding the dynamics of atmospheric circulation in the Northern Hemisphere, especially with regard to the monsoon climate system (deMenocaal and Rind, 1993; Gu et al., 1993; Li, 1990; Prell and Kutzbach, 1987, 1992). The extensive summer heating of the plateau establishes and maintains the Asian summer monsoon circulation (Ren, 1981). In view of the growing concern over a CO_2-enhanced greenhouse effect, there is a pressing need to develop a better understanding of the natural climatic variability on two time scales: decadal and centennial (Sirocko et al., 1993; Singh et al., 1990).

Existing records on glacial/interglacial time scales lack high-temporal resolution. The variation of the last 2000 years is important in providing a useful perspective for several reasons. First, it is a period most relevant to human activities. Secondly, it is a time of identifiable events such as the Little Ice Age and Medieval Warm Period. Thirdly, multi-proxy reconstruction is possible by comparing data from pollen, stable isotopes, tree-rings, and historical records (Thompson, 1996).

However, there are few high-resolution records from the Qinghai-Tibetan Plateau for the last 2000 years. Over the last 2000 years, the high accumulation on ice caps makes it possible to recover pollen records of high temporal resolution, which can serve as proxy indicators for temperature and moisture changes. This chapter focuses on reconstructing climatic changes on the time-scales of decades and centuries from a pollen study of the Dunde ice core. A question of interest is whether the climatic changes in the Qinghai-Tibetan Plateau are linked to large-scale atmospheric circulation that affects a larger area of the globe. In other words, the Dunde ice-core pollen data are used to examine if teleconnection exists between climatic changes in the Qinghai-Tibetan Plateau and other areas of China and the world.

7.2 The Study Region

The Dunde Ice Cap (38°06'E, 96°24'N; 5,325 m) is located on the northeastern margin of the Qinghai-Tibetan Plateau. The Plateau experiences a marked seasonal cycle in which 70 to 80% of the precipitation falls during the summer. The mean annual temperature on the Dunde Ice Cap is -7.3°C (Thompson et al., 1989, 1993; Thompson, 1992). The Dunde Ice Cap is 140 m thick and overlies fairly flat bedrock topography (van Der Veen and Whillans, 1992). It is bordered by deserts of the Qaidam Basin to the south and west and the Gobi Desert to the north. Alpine meadows, steppes and desert, dominated by Cyperaceae, *Artemisia*, and Chenopodiaceae, respectively, occur around the ice cap (K-b Liu et al., 1998). Arboreal vegetation occurs more than 100 km away from the Dunde Ice Cap (Liu, 1997).

7.3 Pollen Data

High annual accumulation rates in the upper part of the Dunde ice core make it possible to undertake high-resolution pollen analysis. For this study, I focus on the last 2000 years of pollen data. The pollen data from four ice cores: deep core 1 and core 3, and shallow core 1 and core 2 (detailed in Chapter III). In order to obtain equal intervals between samples for spectral analysis, pollen data are aggregated into ten-year averages. The laboratory procedure for pollen analysis is briefly described in Chapter III. The relationship between pollen and climate has been discussed in Chapter IV.

7.4 Proxy Data

One advantage of working on ice cores in the Qinghai-Tibetan Plateau is that pollen data can be directly compared with several independent indicators of past climatic changes. These include: (1) proxy data from the same ice core, such as isotopic records, dust, and annual accumulation rate, (2) tree-ring records from Qilian mountain (Zhang and Wu, 1996) and Dulan (Kang et al, 1997) near the Dunde Ice Cap, and (3) historical records (Zhu, 1973; Zhang, 1984; Academy of Meteorology Science, 1983). These paleoclimatic records provide opportunities for the assessment of broad-scale vegetation-climate relationship through time.

7.5 Statistical Methods

In this study, time-series analysis (spectral analysis) was used to estimate the periodicity of climatic changes as reflected in the pollen record, and cross-correlation is used for testing the correlation between pollen and other proxy data. A time series is any set of data ordered through time. The assumptions made in analyzing time series are: (1) all time intervals between adjacent samples are equal; (2) the data are stationary—that is, the series contains no trends in mean or variance. The variance of a time series can be analyzed with respect to either time or frequency. In the time domain, the dependence of values in one series X on past or future values in another series Y is measured by the cross-correlation $p_{xy}(k)$, with time lag k sample intervals. The special case $p_{xy}(k)$ is called the serial or autocorrelation of x at lag k. In the frequency domain the power spectrum of a time series X measures relative contributions $S_x(f)$ to the variance of X from cycles of each frequency f. To minimize standard errors when computing power spectra, "lag windows" are used. Bartlett's window (Platt and Denman, 1975), used here, provides a compromise between reducing standard error of spectral estimates and reducing bandwidth (the range of frequencies over which the window "smudges" spectral peaks). Detailed accounts of time series analysis, including relevant formulae, are given by Platt and Denman (1975).

7.6 Pollen Record

Figure 7.1 shows pollen records from composite cores for the last 2000 years, which suggests several changes in the vegetation. To discuss these changes, seven pollen zones are divided based on both pollen concentration and pollen percentage data.

Zone 1 (A.D. 0-280)

The pollen zone is dominated by *Artemisia*, accompanied with Chenopodiaceae, *Ephedra*, Cyperaceae and some tree pollen. The pollen concentration is relatively high. Percentages of indeterminable pollen, probably an indicator of redeposition by wind in winter, are high.

Zone 2 (A.D. 280-640)

The pollen concentration decreases, and the pollen percentages of Chenopodiaceae are consistently high (>40%). Meadow pollen is absent in this period. It probably represents a dry and cold period. The frequencies of

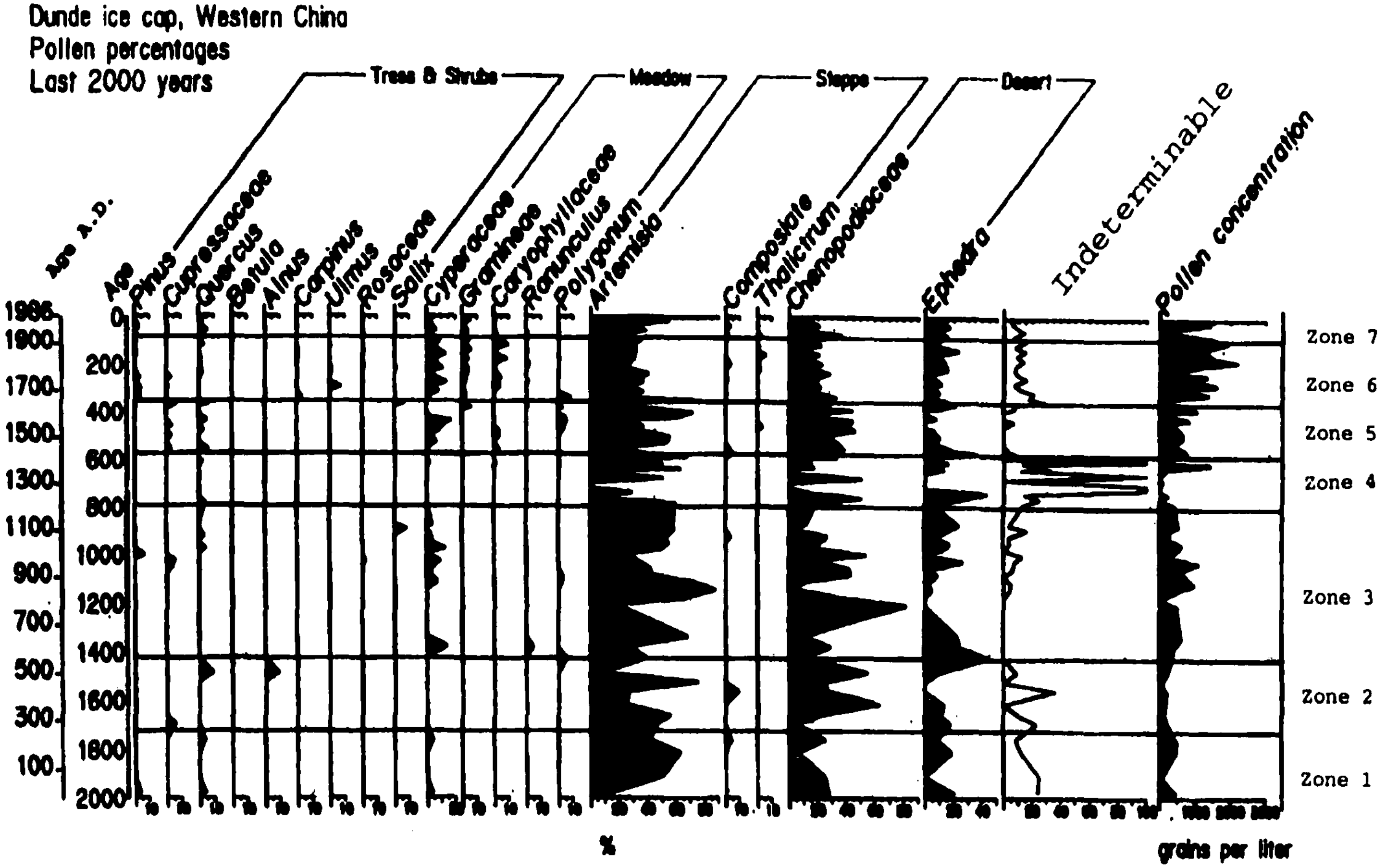


Figure 7.1 Pollen diagram of the Dunde ice core for the last 2000 years. Before 1500 B.P., each sample is 10 year average. Beyond it is 50-year average.

indeterminable pollen are high (>20%), suggesting windy and cold conditions in winter.

Zone 3 (A.D. 640-1200)

Pollen concentrations increase from an average about 246 grains/liter in zone 2 to an average of 521 grains/liter. The pollen zone is characterized by increasing percentages of Cyperaceae and *Artemisia* and very low frequencies of indeterminable pollen. The percentages of *Ephedra* are low. It reflects a warm and humid climate.

Zone 4 (A.D. 1200-1370)

The pollen concentrations are extremely low (average less than 20 grains/liter), accompanied with high percentages of indeterminable pollen. The pollen assemblage is characterized by a decrease in the *Artemisia*, Chenopodiaceae, and *Ephedra*. The low pollen concentrations and lower pollen diversity reflect a low vegetation density, suggesting a warm and dry climate. It probably correlates with the later part of the European "Medieval Warm Period" (Hughes, 1994a; nominally assigned to A.D. 900-1300). The high abundance of indeterminable pollen probably indicates very windy condition.

Zone 5 (A.D. 1370-1650)

The pollen concentration increases to about 500 grains/liter, back to the same level as zone 3. The pollen assemblage was dominated by *Artemisia* (40%) and

Chenopodiaceae (30%), along with Cyperaceae and some tree pollen such as *Quercus* and Cupressaceae. The indeterminable pollen virtually disappears. It suggests an increase in moisture comparing to pollen zone 4.

Zone 6 (A.D. 1650-1926)

The pollen concentration increases to an average of 1132 grains/liter, the highest value over the last 2000 years. Although the pollen assemblage is still dominated by *Artemisia* and Chenopodiaceae, their percentages drop to 30% and 20%, respectively. However, percentages of Gramineae and Caryophyllaceae increase. The increase of indeterminable pollen indicates a cold and windy winter. Therefore, this is a cold and humid condition, probably correlated with the Little Ice Age.

Zone 7 (A.D. 1926-1986)

The pollen concentrations decrease to about 520 grains/liter. Figure 7.2 provides details of the variations of the pollen percentages at annual or biannual resolution from composite cores during the last 100 years and the total pollen concentration using a five-year running mean. It can be divided into two subzones. From 1926 to 1965 (Zone 7a), the pollen concentration decreased to an average of 340 grains/liter. Intitially *Artemisia* (steppe) pollen increased, but it declined subsequently and was replaced by tree and shrub pollen. From 1965 to the present (zone 7b),

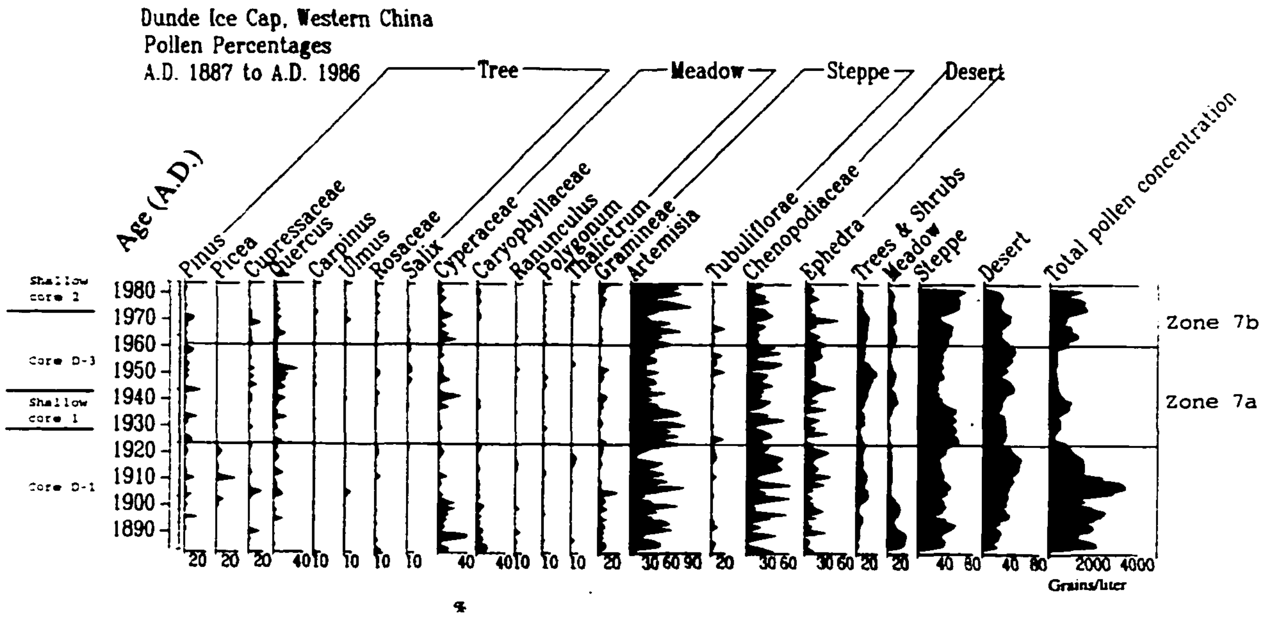


Figure 7.2 Pollen percentage diagram from four cores at Dunde for the last 100 years. The boundaries of cores (left) are not the same as the pollen changes, so it is reasonable to combine them together. The timescale is annual layers back to 1926, and beyond that biannual. The total percentages of tree and shrub, steppe, desert, and total pollen concentration are a 5-year running mean.

the pollen concentration increased to 930 grains/liter, but was still short of the level that it had before the 1920s. This distinct shift after 1965 was accompanied by a notable increase in the percentages of meadow components (Cyperaceae, *Polygonum*) and tree pollen (*Quercus*, Cupressaceae), and a decrease in *Ephedra*, a desert shrub. All these indicate a climatic change towards wetter summers and warmer winters in the Dunde regions (K-b Liu et al., 1998). These effects may be part of a continent-scale summer climatic jump that occurred in the subtropical to mid-latitude regions across the African and Eurasian continents during the mid-1960s (Yan et al., 1990; zhang and Ge, 1990).

7.7 Inferred Climatic Changes from the Ice Core

The pollen record of the Dunde ice core provides the first glimpse at the detailed history of vegetation and climate in northeastern Qinghai-Tibetan Plateau during the last 2000 years. The pollen diagram implies several major climate changes over the last 2000 years.

A humid period during A.D. 0-280 (Pollen zone 1)

Vegetation was characterized by an *Artemisia*-dominated steppe, along with few desert plants, implying a relatively humid period. P.Y. Zhang et al. (1997) studied the historical records of flood/drought events and found two main abrupt climatic changes during the last 2000 years. One of them abruptly changed from frequent flood to drought occurring at A.D. 280. They also pointed out that the most humid climate occurred before 280, and severe drought occurred between A.D. 280 and 500.

A study from a salt-lake in Inner Mongolia showed four main climatic changes occurred ^{14}C age at 1700 BP (A.D. 344), 1500 BP (A.D. 535), 1100 BP (A.D. 910), and 800 BP (A.D. 1256), respectively, during the last 2000 years (Wei, 1992). The lake that changes from brackish to saline occurred at 1700 BP in Inner Mongolia, probably indicating a humid condition before 1700 BP. It may correspond to the vegetation change that occurred at A.D. 280 in the Dunde area.

High percentages of indeterminable pollen indicates cold and windy condition in winter. Historical records from China also showed cold conditions during this period. Historical evidence from Huaihe River showed that it was frozen at A.D. 225 (Zhu, 1973). Zhu (1973) also pointed out that during A.D. 155-220, citrus trees could survive in Xian but could not flower and set fruit. It can be concluded that it was cold and relatively humid between A.D. 0 and A.D.280.

A dry period during A.D. 280-640 (Pollen zone 2)

A steppe-desert dominated by Chenopodiaceae and *Artemisia* co-dominated desert replaced the former *Artemisia*-dominated steppe. Meadow pollen was absent. All these indicated that climate became drier than before. A pollen record from the Qinghai lake (Du et al., 1989) showed that the vegetation abruptly changed from forest-steppe to steppe at 1500 BP (A.D. 535), implying climatic changes from humid to dry conditions. Several Chinese historical records also revealed dry conditions, centered at A.D. 401 (Zhu, 1973). Schove (1949) noted that the large-scale southward migration of people took place in China in the fourth century owing to prolonged dry conditions in north and northwest China. The consequences of the prolonged drought of the fourth and fifth centuries on agriculture and on the population were severe (Zhang and Li, 1986). During this period, in Gansu province, extensive drought resulted in 'extremely high prices of food; one "dou" (approximately 7.5 kg) of grain cost five thousand coins'; 'number of deaths due to hunger estimated to be more than one hundred thousand.' (Gong and Hameed, 1991). The percentage of indeterminable pollen is still high, probably suggesting a cold and windy winter. Therefore, climate probably is dry and cold in the Dunde area during this period.

A warm and relatively humid period between A.D. 640-1200 (Pollen Zone 3)

Vegetation changed from Chenopodiaceae and *Artemisia*-dominated steppe-desert to *Artemisia*-dominated steppe again. The higher pollen concentration suggested that vegetation density was higher. This period is coincident with the Sui and Tang Dynasty Warm Period in China. According to historical records (Zhu, 1973), during this period, there was no snow and ice in Xian (Capital of Tang Dynasty) during winter, and citrus trees grew in this area. However, at present snow falls in Xian in winter, and citrus trees can survive only in places 5 degrees south of Xian in China. So, the record indicates a climate warmer and wetter than present. The tree-ring data from Dulan (Kang et al., 1997), south of Dunde, show a longest warm period from A.D 819 to 1086 in its 1835-year record. It suggests that the early Medieval Warm Period was warm and wet.

A warm and dry period between A.D. 1200-1370 (Pollen zone 4)

The lowest pollen concentrations and higher percentages of indeterminable pollen indicate sparse vegetation cover, suggesting a warm and dry period, which probably corresponds with the late European Medieval Warm Period. Oxygen isotope data from Dunde also suggest a warm period, centered at ca. 700 BP (Yao and Thompson, 1992). Historical records from central China showed that the limit of citrus and ramie (*Boehmeria nivea*) cultivation around A.D. 1264 was at its northernmost, indicating that this was the warmest time during the last 2000 years (Zhang, 1994). Based on the bioclimatic requirements of these cultivated species, it can be estimated that the annual mean temperature in the 13th century was 0.9-1.0°C higher than that at present. In addition, analysis of Chinese flood/drought documentary records revealed a remarkable dry period in the semi-arid regions of west China during A.D. 1100-1500 (Shi, 1985; Gong and Hammed, 1991). During this period the population of four towns (Duenhuang, Wuwei, Jiuquan, Zhangyie) along the Hexi Corridor (part of the Silk Road) in Gansu Province, immediately north and northeast of the Dunde Ice Cap, declined significantly due to drought and famine. Li and Li (1988) reported a 1604-year series of wet/dry indices in Xian. The authors found that between A.D. 1140 and A.D. 1220 it was one of three driest periods (the other two are periods of A.D. 460-500, which probably correlated with zone 2, and A.D. 1420-1490). All this evidence suggests that at least the late Medieval Warm Period was a distinctively warm and dry episode in the Qinghai-Tibetan Plateau.

Hughes (1994) reviewed most of the literatures on the Medieval Warm Event and found that the warmer regional episodes were not strongly synchronous. He also proposed that the high-elevation records are more sensitive to warming than the low-elevation records. Recent work from China (Man and Zhang, 1994) suggests that, although during the 9^{th} to 13^{th} century two short cold periods were evident in east China, the warm period was still pronounced with an average temperature of about 0.5°C higher than present. Therefore, it can be concluded that in the Dunde area the early Medieval Warm Period (9th to 12th century) was warm and humid, while the late Medieval Warm Period (13th to late 14th century) was warm and dry in the Qinghai-Tibetan Plateau.

A slightly humid period between A.D. 1370-1650 (Pollen zone 5)

In this period, vegetation reverted to *Artemisia* and Chenopodiaceae-dominated steppe. Percentages of indeterminable pollen were very low. Some tree and meadow pollen types increased. These indicate that climate changed from a warm and dry condition to a relatively humid condition. Oxygen isotope data from Dunde show a decrease in temperature (Yao and Thompson, 1992). Tree-ring data from Dulan (Kang et al., 1997) showed warm condition during A.D. 1345-1421 and cold conditionduring A.D. 1422-1529, with the cold period lasting longer than the warm period. A decrease in temperature resulted in an increase in effective moisture (Chapter V). This period probably marked a transition from warm to cold conditions.

A cold and wet period between A.D. 1650-1926 (Pollen Zone 6)

The pollen data show that higher pollen concentration occurred during A.D. 1660 to 1926, suggesting a cold and wet period, which could be correlated with the European Little Ice Age cooling during the 17th and 19th centuries (Yao et al., 1991). Oxygen isotope data from the Dunde ice core show that three cold periods and two warm periods occurred during the Little Ice Age (Yao et al., 1991). The pollen data are consistent with the oxygen isotope results. Two lower pollen concentrations during 240-220 BP (A.D. 1740-1760) and 140-120 BP (A.D. 1640-1660) probably corresponded to the two warm periods.

Some authors wrote that the Little Ice Age may have been concentrated in the Northern Hemisphere (Grove, 1988). However, evidence of tropical ice core, from Peru (Thompson, 1992) and lake-level history of tropical African Lake Bosumtwi (Keigwin, 1996; Talbot and Delibrias, 1977) suggest that the Little Ice Age may be a global event.

It is conceivable that climatic changes during the Little Ice Age are different in different regions in the world. Data from the Dunde ice core show generally humid and cold conditions, but with two short warm periods. An ice-core study from Peru showed a humid condition during the early LIA and a dry condition during the late LIA (Thompson, 1992). The lake level history of tropical African Lake Bosumtwi suggests extended drought and lowest lake levels during the Little Ice Age (Keigwin, 1996). The causes of the climatic change during the LIA are not well-understood, but changes in atmospheric trace gases, solar output, volcanic aerosols, and processes internal to the climate system may have played important roles (Grove, 1988).

A warm and dry trend during last 80 years (Pollen zone 7)

The total pollen concentration decreased from an average of 1132 grains/liter before the 1920s to an average of 346 grains/liter afterwards. The meadow pollen also dramatically decreased over that time, implying a warm and dry condition for the last 80 years in the Qinghai-Tibetan Plateau. The oxygen isotope data and annual accumulation rate together also suggest a warm and dry period after the 1920s (Thompson et al., 1993). Evidence from hydrological records of the Yangtze River, the largest river in China, shows drying trend since the 1920s (Fu, 1994).

An eigenvector analysis (Fu, 1994) of the Northern Hemisphere sea-level pressure for June to August shows an abrupt change from positive to negative values in the mid-1920s, indicating an enhancement of the subtropical high over Southeast Asia and the Western Pacific after that time. This atmospheric circulation pattern is favorable for the development of a dry climate in China (Fu, 1994; Sheng, 1986).

In conclusion, the climate in the Dunde area changed from relatively humid (probably cold) during A.D. 0-280 to a dry period during A.D. 280-640. During A.D. 640-1200, the climate became warm and wet, corresponding to the well-documented Sui and Tang Warm Period in China. During A.D. 1200-1370, the climate was warm and dry, probably correlated with the late European MWP. After that, the climate became cold and wet until 1926. During the last 80 years, the climate was warm and dry, probably due to an enhancement of the subtropical high over Southeast Asia and the Western Pacific (Fu, 1994). This circulation pattern may also lead to fewer typhoons affecting China (Sheng, 1986).

7.8 Comparison between Pollen and other Ice-core Proxy Data

Pollen data provided a high-temporal-resolution picture of climatic changes in the Dunde area. An interesting question concerns the relationship between the pollen data and other proxy data from the same ice cores. Two types of proxy data in ice cores have been broadly used in reconstructing palaeoclimate: oxygen isotope and dust data. In this study, I will compare total pollen concentrations with oxygen isotope and dust concentrations within the same ice core.

$\delta^{18}O$ is widely regarded as a proxy for temperature (Dansgaard et al., 1969; Jouzel et al., 1994; Thompson et al., 1989, 1997). Yao et al. (1996) suggested that oxygen isotope value is positively correlated with air temperature on the Qinghai-Tibetan Plateau. Therefore, I hypothesize that oxygen isotope values should be negatively correlated with total pollen concentrations within the same core based on our findings that total pollen concentration is negatively correlated with temperature at Dunde. In order to test this hypothesis, a 1500-year stratigraphic sequence of pollen and oxygen isotopes at 10-year average is used. Figure 7.3 presents the relationship between total pollen concentration and oxygen isotope data within the same ice core. There is no correlation between total pollen concentration and oxygen isotope data at the 95% significance level. As we know, the oxygen isotopic value is considered to be an indicator of temperature. Yet, local isotopic values may change as a result of factors other than temperature (Charles et al., 1994, 1995). The seasonal variation in the timing of precipitation events may also affect the oxygen isotope values, regardless of the source region (Charles et al., 1995). On the other hand, oxygen isotope can be significantly depressed by an increase in precipitation in part of the Qinghai-Tibetan Plateau due to intensified summer monsoon (K-b Liu et al., 1998; Yao et al., 1996).

Is there other relationship between total pollen concentration and oxygen isotopic values? Cross-correlation analysis (Figure 7.4a) shows that total pollen concentrations are negatively correlated with $\delta^{18}O$ values at the zero point (no sample shift), and are positively correlated with $\delta^{18}O$ values when total pollen concentration is 2 samples ahead. This result probably indicates that oxygen isotope and total pollen concentration come from different sources. Seasonal

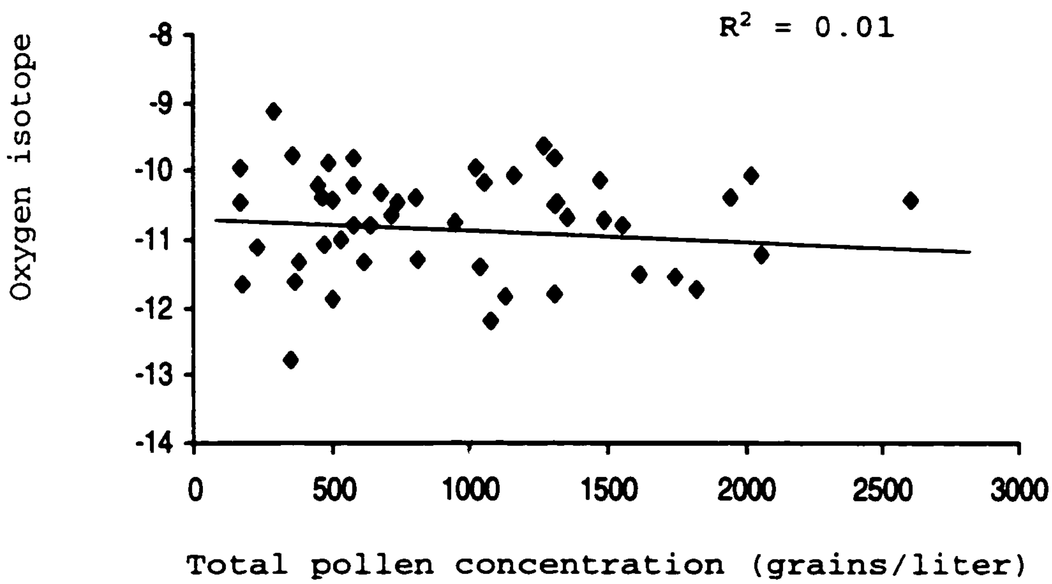
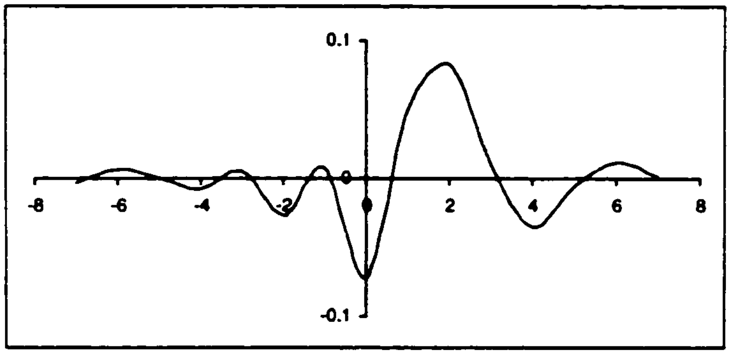
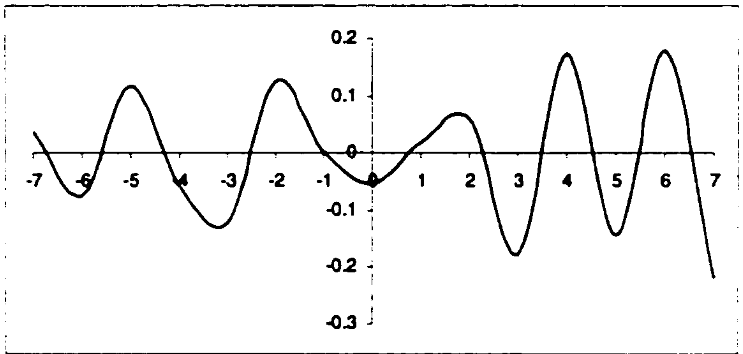


Figure 7.3 The relationship between pollen concentration and oxygen isotope data from core D-3 in the Dunde Ice Cap. For a 1500-year record. Solid line is regression line.



(a)



(b)

Figure 7.4 The results of cross-correlation analysis between total pollen concentration and two variables: (a) oxygen isotope, (b) big particle data. X-axis = number of lag, y-axis = correlation coefficient.

pollen studies show that significant amounts of pollen are found in winter samples due to redeposition after the flowing season (Liu, 1997; Van Campo et al., 1996). Therefore, the relationship between total pollen concentration and oxygen isotope data should be non-linear and more complicated.

The dust particles in the Dunde ice core come from areas west of Dunde (Thompson et al., 1989; Liu et al., 1999). Dust concentrations can be used as an indicator of atmospheric circulation and turbidity (Mosley-Thompson and Thompson, 1994; Mosley-Thompson et al., 1993, 1995). One question is whether the total pollen concentration is correlated with dust data. The result of cross-correlation analysis (Figure 7.4b) show that pollen concentrations are negatively correlated with dust particle concentrations at the zero point (no sample shift). However, the correlation coefficient increases after samples of pollen concentrations are ahead or lag that of dust concentrations, probably indicating that the relationship between total pollen concentration and dust data is more complicated. The reason is that the pollen and dust come from different regions and transported by different ways. On the other hand, high dustfall frequencies (higher dust concentration) usually occur in a cold and dry environment (Zhang, 1984), while high pollen concentration occur at cold and humid conditions (K-b Liu et al., 1998). Therefore, it is reasonable that there is not significant linear relationship between total pollen concentration and dust data.

7.9 Discussions

7.9.1 Climatic Change in the Qinghai-Tibetan Plateau during the Last 2000 Years

Pollen records from the Dunde ice core reveal a detailed history of climatic changes over the last 2000 years. It is comparable to other climatic proxy records retrieved from the Qinghai-Tibetan Plateau and surrounding areas. Figure 7.5 presents a comparison of three high-resolution records: pollen concentration record from the Dunde ice core, net accumulation data from the Guliya ice core, and moisture index curve from the semi-arid region of western China. It is reasonable to compare them, because they are indicators of moisture variability. The three data sets show a comparable climatic history.

Five major climatic periods delineated from the pollen concentration data from Dunde are basically identical with those recorded in the net accumulation data set from the Guliya ice core (Figure 7.5), although the beginning or ending times for these periods are different. The difference in timing between Dunde and Guliya may reflect spatial variability in moisture source over the plateau. Observations of contemporary climate regimes on the plateau reveal such spatial difference (Luo and Yanai, 1983, 1984; Tang et al., 1979) (Figure 7.6). A comparison of the two $\delta^{18}O$

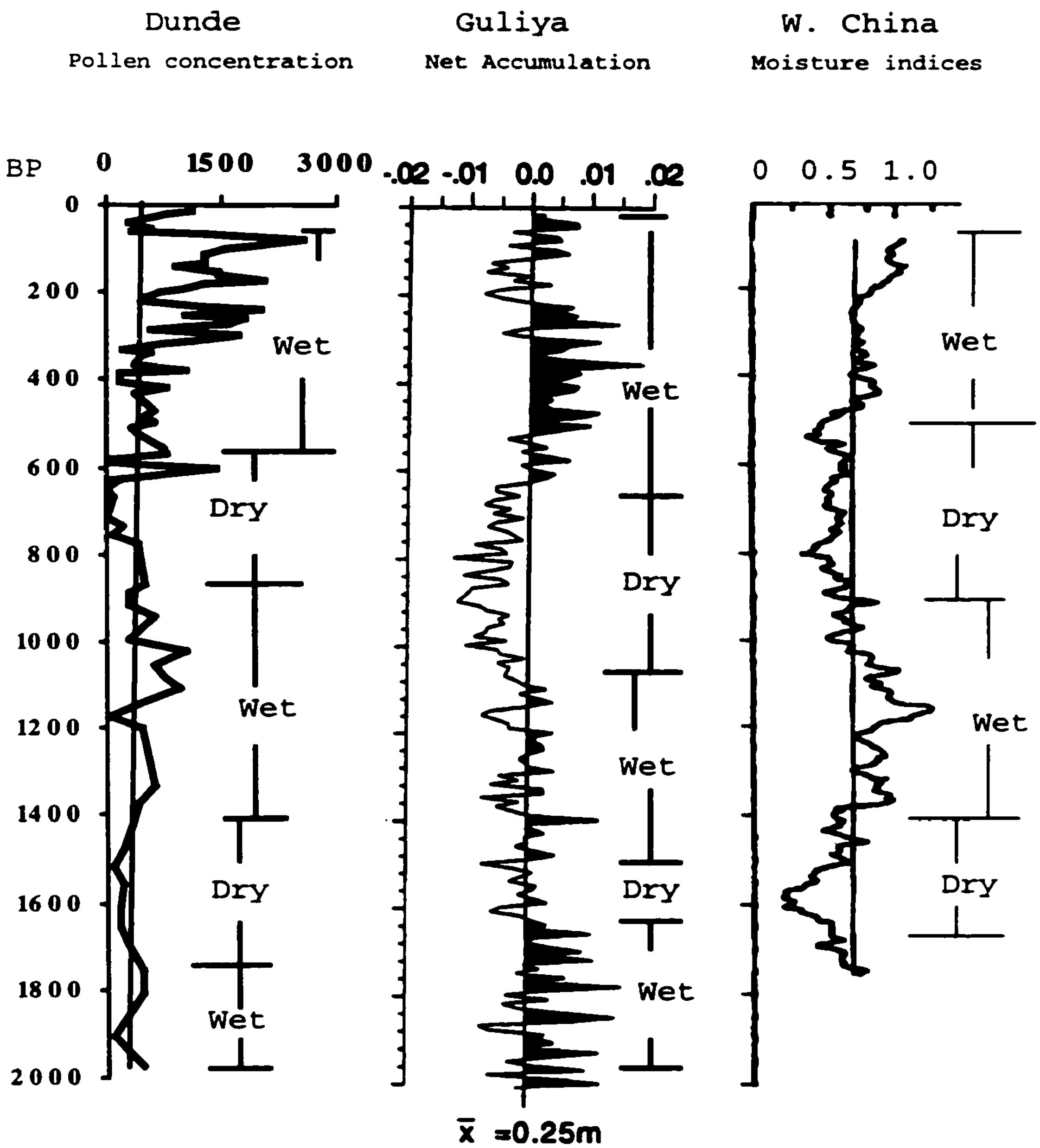


Figure 7.5 The 2000-year records of pollen concentration as decadal average on the Dunde compared with net accumulation on the Guliya ice cap (Thompson, 1996), and variation of the moisture indices for semi-arid regions of western China. (After Gong and Hameed, 1991).

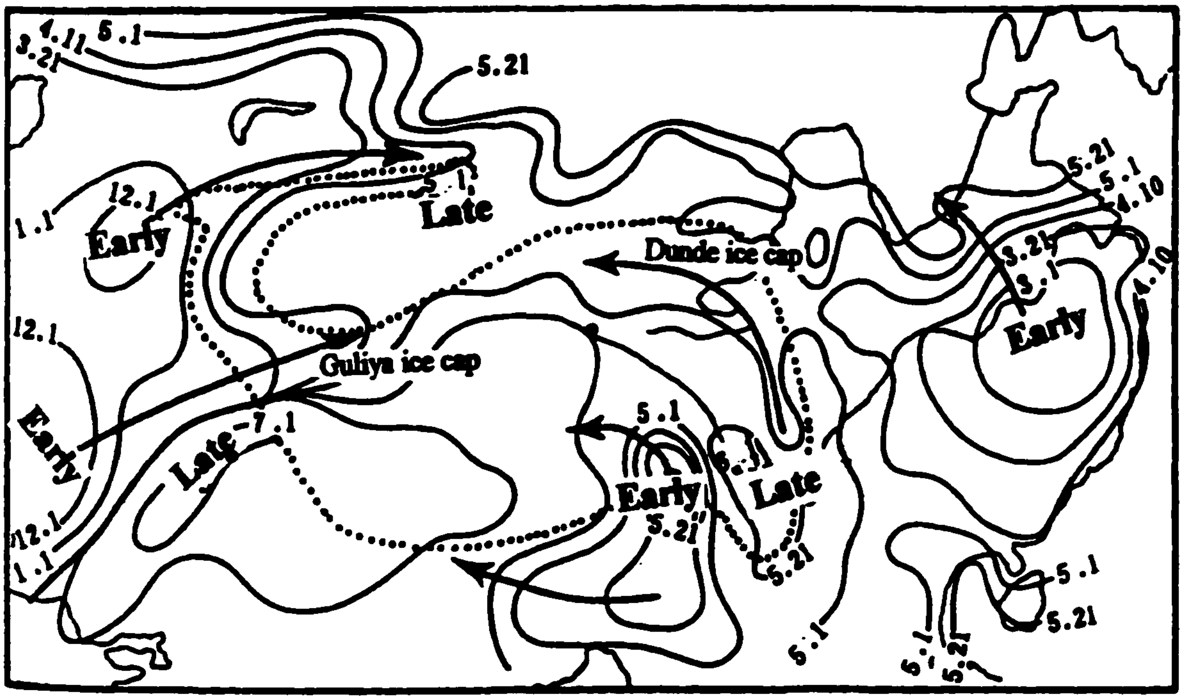


Figure 7.6 The contours of beginning rain date. Arrow represents the direction of rain zone moving. The dashed line represents the contours of elevation equal 3000 m. (Modified from Tang et al., 1979)

histories from the Dunde and Guliya ice cores for the last 1,000 years suggests that the temperature trends on these two ice caps also were dissimilar at times and often in anti-phase (Figure 2.22).

The moisture record from the Dunde ice core shows a broad similarity to that from the semi-arid area in western China (Figure 7.5). The two wet and two dry periods that occurred in western China can be correlated with those recorded in the pollen concentration dataset in Dunde. It is reasonable because the two areas are located in the steppe vegetation region, under the same climatic system affected by the summer and winter monsoons.

In addition, the Dunde pollen records show that two cold periods occurred in A.D. 0-640 and A.D. 1650-1910. They are not consistent with those identified from the oxygen isotope records in the same ice core. The oxygen isotope data using 100-year averages from Dunde show two cold periods centered at A.D. 1700 and A.D. 800 (Yao and Thompson, 1992; Yao et al., 1991). However, the temperature changes inferred from the Dunde pollen record are comparable to other proxy data in the plateau. Recently, high-resolution temperature records, based on changes of magnetic susceptibility, sedimentary pigments, and stable carbon isotope ($\delta^{13}C_{org}$) values, are recovered from Lake Ximencuo (33°23'N, 101°08'E) in eastern Qinhghai-Tibetan Plateau for the last 2000 years (Wang et al., 1997). Two cold periods are identified at A.D. 0-480, and A.D. 1460-1900, correlated with two cold periods from Dunde. In addition, according to historical records about cold events in Tibet, Wu (1990) found two pronounced cold periods that occurred at A.D. 0 to 500 and A.D. 1500-1900. Tree-ring data from Dulan also provide a detailed temperature record (Kang et al., 1997). The authors distinguished 11 warm periods and 10 cold periods. The longest warm period occurred between A.D. 819 and A.D. 1086 that may correlate with pollen zone 3 in Dunde. Three coldest periods in Dulan during A.D. 1422-1833 probably correlate with pollen zone 6 in Dunde.

Therefore, although the chronologies of climatic change inferred from different proxy datasets are different, at least three wet periods and two cold periods occurred over the Qinghai-Tibetan Plateau during the last 2000 years. One of the two cold periods occurred early in the first millennium and the other in the later part of the second millennium.

7.9.2 Periodicity of Climatic Change in the Dunde Ice Core

Spectral analysis was performed to reveal the periodicity of climatic changes at Dunde. Figure 7.7 shows the result of spectral analysis, and the pronounced cycles are 88 and 40 yr for total pollen concentration, 200, 110

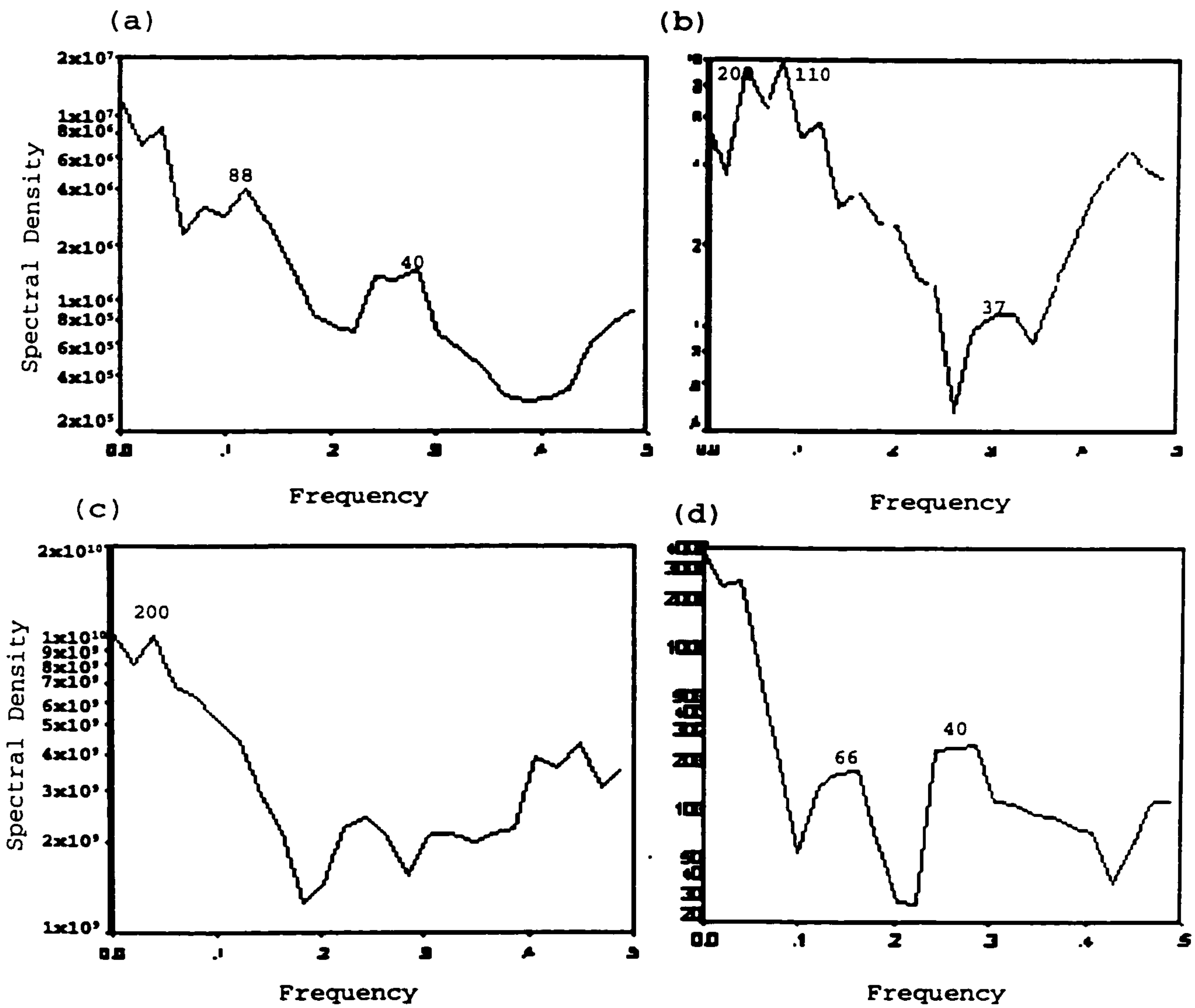


Figure 7.7 The result of spectral analysis from the Dunde ice core. (a) total pollen concentration, (b) oxygen isotope (c) big dust concentration, (d) electrical conductivity. The x-axis = frequency y axis = density.

and 37 yr for oxygen isotope, 200 and 40 yr for big dust concentrations, and 66 and 40 yr for electrical conductivity. The cycles for these variables are different.

Stocker and Mysak (1992) reviewed the century time scale climatic variability that is observed in high-resolution proxy records derived from ice cores, tree-ring index series, pollen records, and sea ice. The review has clearly shown that climate fluctuations occurring on the century time scale, i.e. 50-400 yr, are a ubiquitous phenomenon during the Holocene (Figure 7.8). The traditional interpretation is that decadal-to-century scale fluctuations in the climatic system are externally forced, e.g. by variations in solar properties, such as the Gleissberge cycle of 84 yr. Recently, several models of different complexity have been developed to study the thermohaline circulation of the ocean. Deep water is formed in the North Atlantic, from where it flows as deep current into the Pacific to upwell, producing a global conveyor belt circulation. The spectral analysis of the Atlantic heat flux showed two strong cycles of 110 and 38 yr (Stocker and Mysak, 1992). The ocean has classically been considered as a major reservoir of heat and water. The factor that causes the century-scale climatic changes is probably the ocean.

The pollen concentration in Dunde is strongly controlled by summer temperature (Chapter V). An 88-year

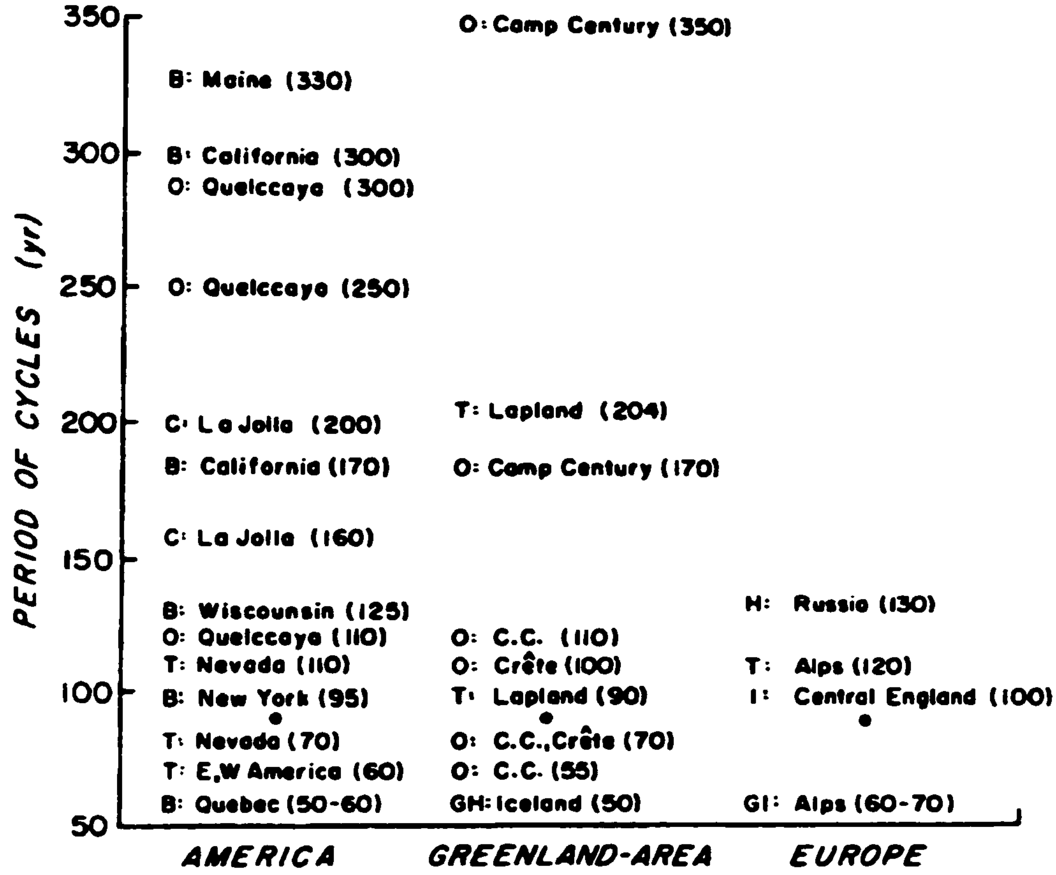


Figure 7.8 Summary of climatic variations observed on the century time scale. The findings are spatial and temporally ordered. The capital letters denote evidence in biological (B), radio carbon (C), historical (H), instrumental (I), oxygen isotope and other ice core parameters (O) and tree-ring records (T). The observed cycle period is given in brackets. C.C. denotes the Camp Century oxygen isotope record and * indicates the 83-yr cycle in marine air and sea surface temperatures of Folland et al., (1984) (After Stocker and Mysak, 1992).

cycle of pollen data is probably correlated with solar activity, such as the Gleissberge cycle of 84 years. However, the oxygen isotopes come from the ocean (Thompson et al., 1989; Aizen et al., 1997), and the 110- and 40-year cycle in the oxygen isotope date is probably correlated with the ocean activity. Therefore, it may be concluded that the different cycles recognized different datasets probably reflect responses to different climatic forcing.

7.10 Conclusions

Pollen data from the Dunde ice core for A.D. 0-280 show relatively high total pollen concentration and high percentages of *Artemisia*, suggesting a cold and wet period. Low pollen concentrations and high percentages of indeterminable pollen suggest a dry and cold period between A.D. 280-640. Following this period, an increase in total pollen concentration and percentages of Cyperaceae pollen indicated a warm and wet period, correlated with the Sui and Tang Warm Period in East China. Between A.D. 1200 and 1370, the lowest pollen concentrations and highest percentages of indeterminable pollen suggest a warm and dry climate, probably correlated with the late European Medieval Warm Period. The highest pollen concentration occurred during 1660-1910 A.D., implying the prevalence of cool, humid conditions during the Little Ice Age. Low pollen concentration and few pollen types indicate a dry climate for the last 80 years.

The three wet and two dry periods inferred from the Dunde pollen record are consistent with records from Guliya (accumulation) and western China (moisture index), suggesting that pollen is a useful indicator of climatic changes, especially as an indicator of moisture availability in Dunde.

Traditional explanation of century-scale climatic change evokes the notion of solar variability. The climatic changes inferred from the Dunde ice core show several pronounced cycles, which possibly suggest that different proxies respond to different forcing mechanisms.

CHAPTER VIII

A HOLOCENE POLLEN AND CHARCOAL STUDY FROM THE GULIYA ICE CAP

8.1 Introduction

Reconstructions of paleoclimates based upon general circulation models (GCMs) are critical to our understanding of late Quaternary events over wide geographic areas. GCM simulations (e.g., Kutzbach, 1981; Kutzbach and Gutter, 1986; COHMAP Members, 1988) provide snapshots of paleoclimates for 3000-year intervals for much of the Northern Hemisphere. Boundary conditions for these simulations infer maximum seasonality (8% greater summer and 8% less winter insolation than today) at about 9000 BP. This led to wetter climatic conditions during the early to middle Holocene in the Qinghai-Tibetan Plateau than exist today (Winkler and Wang, 1993).

The sensitivity and rapid response of ice caps to climatic change makes them a potentially excellent source of information on small-magnitude, short-duration climate changes of the kind that appears to have characterized the late-Quaternary (Thompson et al. 1995a). Paleo-environmental data from the Qinghai-Tibetan Plateau are very important for understanding patterns of past and present global climate circulation. Pollen analysis is one method that provides a view on vegetation dynamics and climate history (K-b Liu et al. 1998). Unfortunately, in high-latitude areas there are usually few opportunities to use palynology for investigating deposits that have continuously accumulated during more than a few millennia. However, ice cores from the low to middle latitudes provide a unique opportunity for palynological investigation of long-term records in parallel with other proxy data from ice cores.

A number of papers are devoted to palynological investigations of ice cores (Fredskild and Wagner, 1974; Lichti-Fedorovich, 1974, 1975a, 1975b; McAndrews, 1984; Short and Holdsworth, 1985; Barry et al, 1981; Bourgeois, 1986; Bourgeois et al., 1985; Koerner et al., 1988; Koerner and Fisher, 1990; Andreev et al., 1997; K-b Liu et al., 1998). These studies have demonstrated the possibility of correlation between pollen fluctuations and climatic events. However, most of these works focus on the polar areas.

This chapter presents a pollen record from the Guliya Ice Cap of northwestern Qinghai-Tibetan Plateau in conjunction with charcoal analysis. These new results allow us to infer past atmospheric circulation patterns and climatic changes and to discuss the causes for climatic changes. The correlation between the Guliya ice-core pollen assemblages with regional pollen diagrams is also discussed.

8.2 The Study Region

The Guliya Ice Cap (35°17'N, 81°29'E; 6710 m.s.a.l) is located in the northwestern corner of the Qinghai-Tibetan Plateau (Figure 8.1). It is the largest (>200 km^2) and thickest (308.6 meters) subtropical ice cap yet investigated (Thompson et al., 1997). It is part of an ice mass that extends over 8000 km^2 in the western Kunlun mountains. In 1992 two deep ice cores were recovered and the longer core (C-2; 308.6 m) reached bedrock. No hiatus was observed in the core, and the visible layers were horizontal throughout. The depth-age model is based on Thompson et al. (1997).

Western Tibet is a cold desert (Arakawa, 1969), where evaporation/precipitation (E/P) ratios range between 20 and 50. The prevailing moisture source for the Guliya Ice Cap is the Arabian Sea (Ohata et al., 1989; Tang et al., 1979). On the Guliya Ice Cap the annual precipitation is low, averaging 180 mm w.e. (Thompson et al., 1995a), compared with 350 mm w.e. at Dunde (Thompson et al., 1989). The region is characterized by low temperatures, particularly during the 3-month growing period. The monthly mean temperature from June to September is less than 15°C (Wang, 1988).

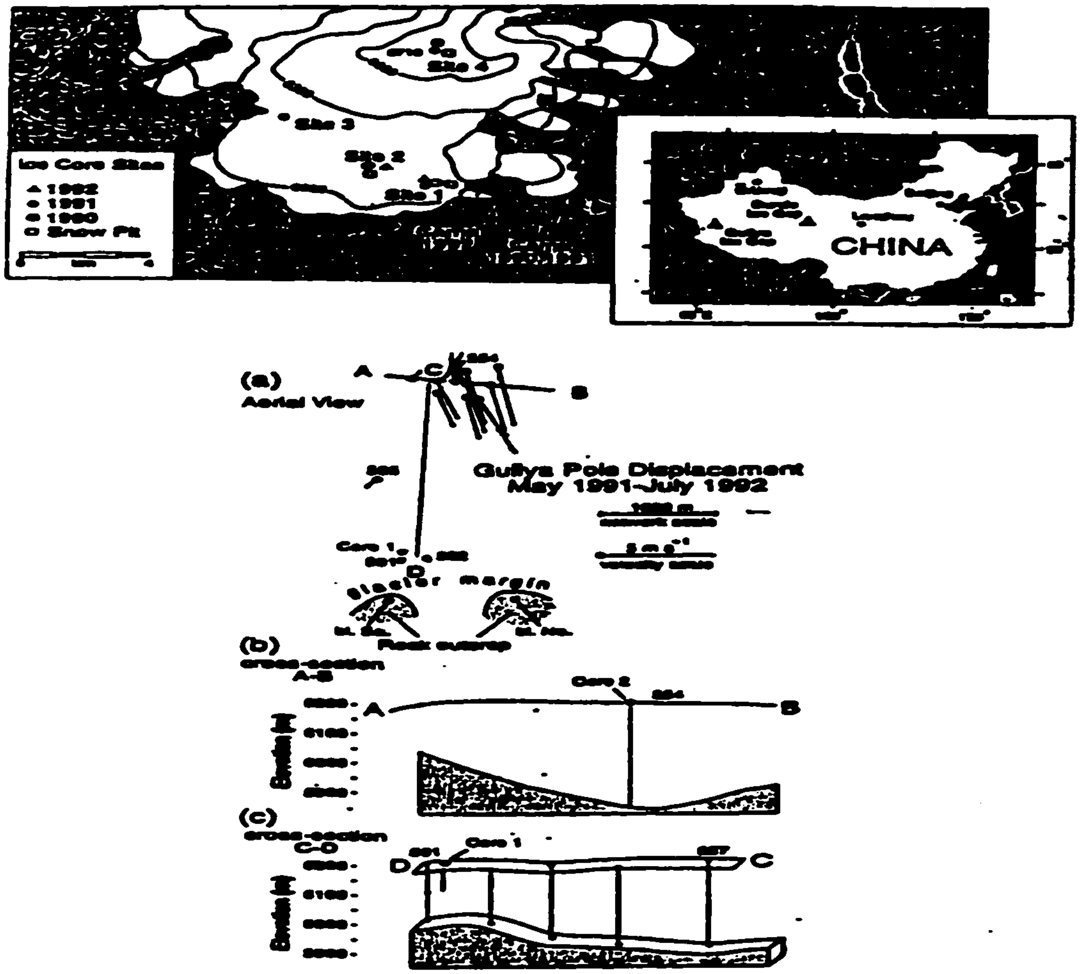


Figure 8.1 Locations of the Guliya and Dunde ice caps and snow-pit and ice-coring sites on Guliya. (a) Aerial view of displacements from May 1991 to July 1992 for core 1 and core 2 drill sites. The ice-thickness profiles were measured during a short-plus radar survey. (b) Transect A-B through the core 2 drill site. (C) Transect C-D from the core 2 drill site to the core 1 drill site. (After Thompson et al., 1997).

Due to the regional difference of temperature and moisture conditions on the Tibetan Plateau, zones of forest, steppe, and desert appear in succession from the southern slopes of the Middle Himalayas in the south to the central part of western Tibet (Chapter III, Figure 3.5). The floristic components of the vegetation include more than 3000 species, mainly from such families as Gramineae, Compositae, and Cyperaceae (Wang, 1981). Other families such as Rosaceae, Cruciferae, Boraginaceae, and Labiateae are well represented. The most important genera are *Stipa* and *Artemisia* (Wang, 1981).

8.3 Methods of Palynological and Charcoal Studies

Pollen

Ice cores retrieved from the Guliya Ice Cap have already produced a wealth of high-temporal-resolution paleoclimatic data for the northwestern Qinghai-Tibetan Plateau (Thompson et al., 1997). Forty-one meltwater samples were used for pollen analysis from a long core, D-1. These meltwater samples ranged in volume from 250-850 ml (mean = 650 mL). From present to 9000 BP, each sample represents a 50-year aggregate, and the interval between adjacent samples is 500 years. Beyond 9000 BP, each sample represented about 100 years. The last sample from Core 2 covers from 14,800 to 15,900 BP.

The pollen processing procedure is the same as that described by K-b Liu et al. (1998). On average, about 80 pollen grains (ranging from 20 to 95 grains) were counted in each sample, and a total of thirty-eight pollen types was identified in all samples. The pollen results are given as a percentage pollen diagram of selected taxa, the percentages being based on the total pollen sum.

Charcoal

Charcoal is an amorphous inorganic carbon compound, which results from the incomplete combustion of plant tissues. This combustion is supported by the generation of the combustible gas carbon monoxide, methane, and low levels of oxygen (Cope and Chaloner, 1980, 1981; Clark and Russell, 1981). Charcoal preserves well and therefore can be used as a record of past fires (Komarek, 1974).

It has been demonstrated that charcoal particles deposited in lake sediments can reveal the history of fire around the lake (Chen, 1990; Patterson et al. 1987; Clark and Russell, 1981; Horn, 1989; Horn and Sanford, 1992; Mworia-Maitima, 1997). Earlier work (Patterson et al., 1987) has shown that in lake sediments charcoal deposits originating from natural fires tend to show short-term peaks of high deposition rates as opposed to the relatively longer period of continuous high deposition in the case of fires from an anthropogenic origin.

Charcoal preserved in lake sediments, peat, and soils provides a record of past fire occurrence (Clark and Hussey, 1996; Clark and Russell, 1981). Unlike pollen, which is produced continuously in fairly constant amounts, charcoal is produced in large quantities but at irregular intervals. These are a function of the fire regimes that are often unique to specific vegetation types and climatic regions. Therefore, I assumed that the change of charcoal particles could be due to more or less vegetation cover around the Guliya Ice Cap.

Charcoal concentration was used in the charcoal analysis (Stockmarr, 1971). Palynological procedures do not significantly alter charcoal size or abundance (Clark, 1984). Charcoal concentration (number of charcoal particles per liter) was calculated using the following formula:

$$C = \frac{N * T}{L * V}$$

C charcoal concentration (number per liter)

N number of charcoal counted in a sample

L counting number of *Lycopodium*

T total number of *Lycopodium* added to a sample

V volume of meltwater

During charcoal counting, the shape and size of charcoal fragments were also noted, although these parameters were not quantified in this study. Umbanhowar and McGrath (1998) have noted that grass charcoal (125-um screen) was significantly more elongate (mean L : W = 3.62) than either leaf (1.91) or wood (2.13) charcoal (Figure 8.2). Most of the charcoal particles from this ice core were more elongated, indicating that they come from grass.

8.4 Ice-core Pollen and Vegetation

The few published surface pollen spectra from the Karakorum and Kunlun Mountains (Weng, 1989; Weng et al., 1993; Van Campo et al., 1996) suggest that most pollen come from the local vegetation. One surface pollen spectrum obtained at 5100 m close to the Guliya ice core contained twenty-five pollen types, among which only 6 equaled or exceeded 1 %: *Artemisia* (46.4%), Chenopodiaceae (29.3%), Gramineae (10.2%), *Ephedra* (3%), Cyperaceae (2%), and Compositae (1.2%) (Van Campo et al., 1996). All these pollen types are from local and regional vegetation. Arboreal pollen, represented by *Salix*, *Alnus*, *Cedrus*, and *Pinus*, accounted for 1.8% of the spectrum. Atmospheric pollen content was studied for one annual cycle from September 1989 to October 1990 on the Kunlun mountain at Tien Shui Hai station (35°40′N, 79°30′E; 4850 m a.s.l) by Van Campo et al. (1996). The most abundant pollen taxa were Chenopodiaceae (33%), *Artemisia* (32.3%), Gramineae (7.7%), Cupressaceae (5.2%), *Ephedra* (4.2%), *Tamarix* (1.8%), Urticaceae (1.7%), Cyperaceae (1.2%), and *Pinus* (0.6%).

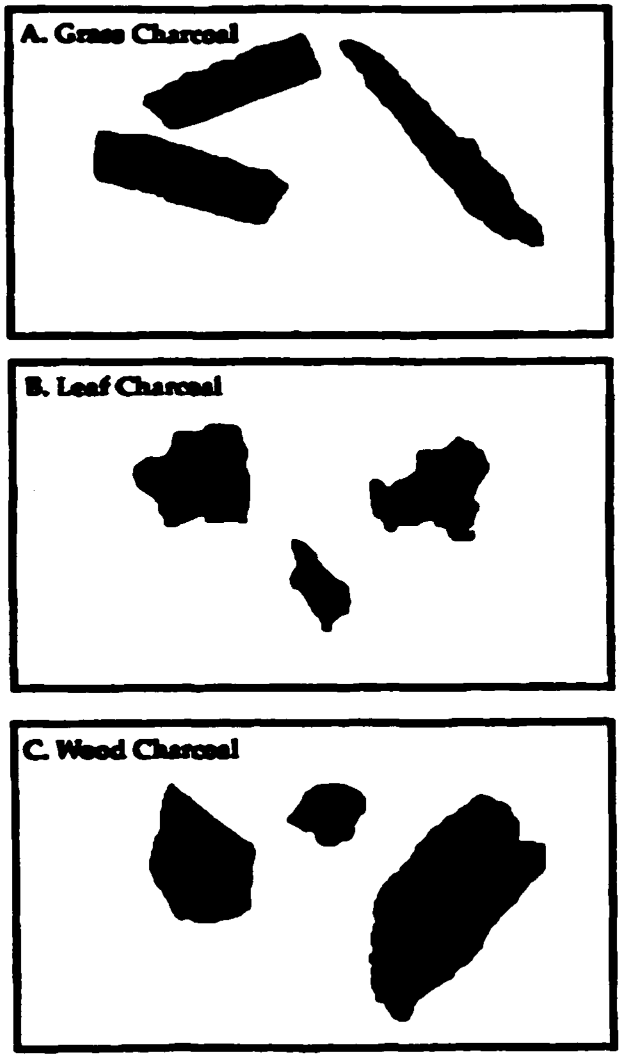


Figure 8.2 Silhouettes of charcoal. (A) grass, (B) leaf, and (C) wood. (After Umbanhowar and McGrath, 1998).

Thirty-one surface sediment samples were collected from West Kunlun Mountains in northwestern China (Weng et al., 1993). The predominant pollen types are Chenopodiaceae, *Artemisia*, *Picea*, *Ephedra*, Gramineae, and Cyperaceae. Regression analysis was applied to reveal the relationship between pollen and plant percentages of *Artemisia*, Chenopodiaceae, Cyperaceae, and Gramineae. The results indicated that a linear relationship existed between pollen and plant percentages for *Artemisia*, Chenopodiaceae and Cyperaceae (Weng et al., 1993)

Thus, it is reasonable to assume that the majority of pollen identified in the Guliya ice core is derived from local vegetation. The presence of exotic tree pollen shows the possibility of long-distance transport. Long-distance pollen transport normally does not significantly influence the pollen spectra. It can be assumed that it was an insignificant component in the past as it is in the present. When they are present, the long-distance transported pollen may provide information about wind direction (Nichols et al., 1978; McAndrews, 1984; Bourgeois, 1986).

8.5 Results of Palynological and Charcoal Studies

Figure 8.3 summarizes the pollen results from the Guliya ice core. Most of pollen types come from steppe and desert vegetation. The tree pollen is derived from long-distance transport. Six main pollen zones can be distinguished.

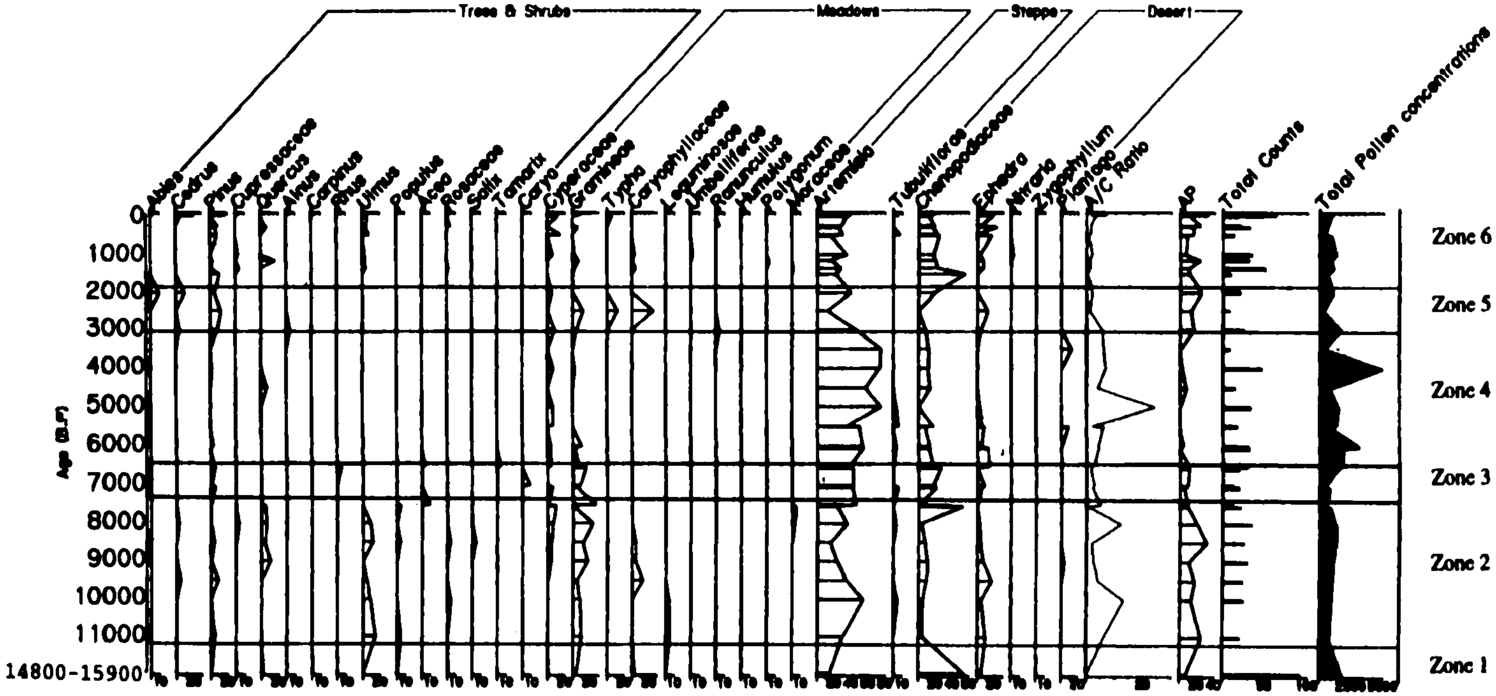


Figure 8.3 Pollen record from the Guliya ice core.

Zone 1 (15,900 to 11,500 BP)

The pollen assemblage is dominated by Chenopodiaceae (60%) and *Artemisia* (about 10%), along with *Ephedra*. The A/C ratio is low. These imply dry conditions during this period.

Zone 2 (11,500 to 9500 BP)

In this zone, the pollen assemblage is dominated by *Artemisia* (30%), Chenopodiaceae (15%), *Ephedra* (10%), and *Ulmus* (10%). It is characterized by a decrease of Chenopodiaceae (from 60% to 15%) and an increase of tree pollen. The A/C ratio is higher than in zone 1. These suggest warm and relatively humid conditions, probably due to onset of an intensified summer monsoon (Van Campo and Gasse, 1993).

Zone 3 (9500 to 6500 BP)

This zone is characterized by an increase of Gramineae and Cyperaceae pollen, although the pollen assemblages are still dominated by *Artemisia* and Chenopodiaceae. However, the A/C ratio decreases. The A/C ratio is less significant mainly because four pollen taxa dominated the pollen spectra. The higher frequencies of Gramineae pollen suggest a wetter climate in the Qinghai-Tibetan Plateau (Chapter IV). Therefore, an expansion of grassland (Gramineae) over *Artemisia*-dominated steppe implies an increase of precipitation during this period. An increase of long distance transported tree pollen such as *Cedrus* and *Quercus* may suggest strong southerly winds from the Himalayas.

Charcoal concentrations also increase during this period (Figure 8.4), accompanied by an increase of Gramineae pollen. This pattern of stratigraphic changes can be interpreted to indicate a period of more frequent wildfires in a grass-dominated steppe environment. It is likely that compared with desert vegetation, grassland ecosystems were more susceptible to wildfires because the vegetation cover was denser and more continuous than in the desert, thus providing more fuel for fires to ignite and spread. The increase in vegetation density and the expansion of steppe during the early to middle Holocene was probably due to higher summer precipitation caused by an intensified summer monsoon, as predicated by the results of climate modeling (Winkler and Wang, 1993).

Zone 4 (6500 to 2900 BP)

This zone is characterized by higher pollen concentrations, an increase of *Artemisia*, a high A/C ratio, and little or no tree pollen. The pollen concentration increases to a maximum at about 4500 BP. Gramineae pollen decreased and became virtually absent, while Cyperaceae pollen increased. These suggest a period of reduced precipitation, resulting in the expansion of *Artemisia*-dominated steppe over grassland. The increase in total

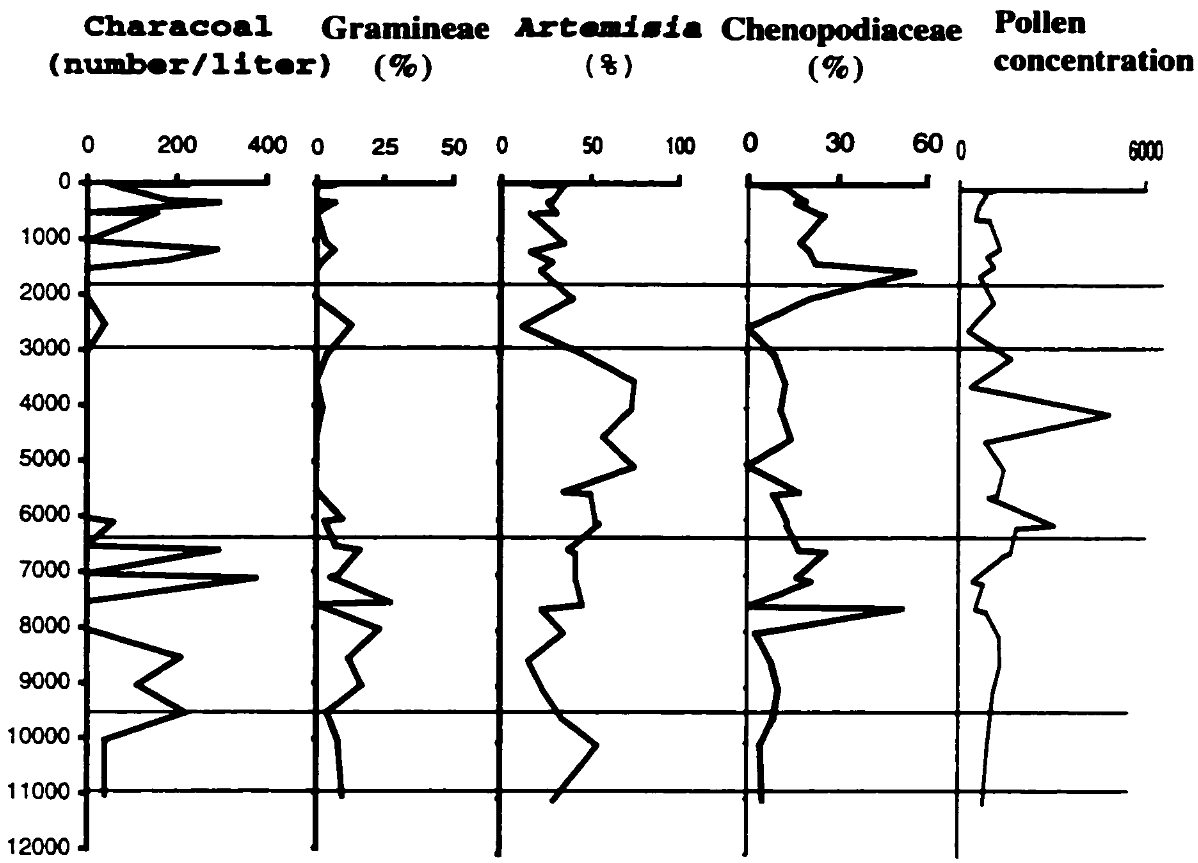


Figure 8.4 Comparison between concentration of charcoal particle and main pollen types in the Guliya ice core.

pollen concentration may be explained by higher pollen production by *Artemisia* than that by Gramineae. The absence of charcoal in the zone is intriguing. A possible explanation is that *Artemisia*-dominated steppes would be less susceptible to the spread of wildfires than would grasslands. The lack of tree pollen may suggest reduced advection of southerly wind from the Himalayas.

Zone 5 (2900 BP to 1900 BP)

This zone is characterized by increased percentages of Gramineae, Caryophyllaceae, *Typha*, and *Pinus*, while percentage of *Artemisia* decreases. *Cedrus* and *Pinus* appear again, and *Abies* pollen appears for the first time. Charcoal concentration increases slightly. All these indicate a relative increase in effective moisture and/or reduction in summer temperature.

Zone 6 (1900 BP to present)

The pollen spectra are characterized by a distinct increase in desert components such as Chenopodiaceae and *Ephedra*, suggesting an increase of aridity. This pollen assemblage is similar to the modern pollen records from Tianshui Hai (Van Campo et al. 1996). This indicates that modern dry climatic conditions in the Guliya area have been established since then.

In summary, pollen records from the Guliya ice core display several changes in the vegetation for late-Quaternary (Table 8.1). From 11500 to 9500 BP, an *Artemisia*-dominated steppe replaced the late-glacial Chenopodiaceae-dominated desert, suggesting an increase of precipitation. Between 9500 BP and 6500 BP, a grass-dominated steppe replaced the *Artemisia*-dominated steppe, implying an increase of precipitation caused by an intensified summer monsoon, as predicted by the results of climate modeling (Winkler and Wang, 1993). From 6500 BP to 2900 BP, vegetation reverted to *Artemisia*-dominated steppe, suggesting a decrease of precipitation. From 2900 to 1900 BP, an increase in the percentages of Gramineae, Caryophyllaceae, and *Typha* indicated a relative increase of effective moisture. Since 1900 BP, the pollen assemblage has been similar to the modern pollen spectra, indicating that modern climatic conditions in the Guliya have been established since then.

8.6 Comparison between the Guliya and Dunde Pollen Records

The Dunde Ice Cap and the Guliya Ice Cap are situated 1400 km apart on opposite sides of the Qinghai-Tibetan Plateau. The Dunde Ice Cap is located on the northeastern margin of the plateau, while Guliya is on the northwestern side (Figure 3.1). A detailed Holocene pollen record from the Dunde ice core has been reported by K-b Liu et al.

Table 8.1 Summary of the climatic changes inferred from the Guliya ice core.

Time	Vegetation	Climate
1900-present	Chenopodiaceae and *Ephedra*-dominated desert	dry
2900-1900 BP	*Artemisia* and Chenopodiaceae-dominated steppe	cold humid
6500-2900 BP	*Artemisia*-dominated steppe	humid
9500-6500 BP	Grass-dominated steppe	most humid
11500-9500 BP	*Artemisia*-dominated steppe	humid
Before 11500 BP	Chenopodiaceae-dominated desert	dry

(1998). The question is whether the two ice-core pollen records have features in common and whether the climatic history reconstructed from the two ice cores is comparable.

Pollen diagrams from these two ice caps have features in common when comparing the major pollen types from the two ice cores. The pollen assemblages are dominated by *Artemisia*, Chenopodiaceae, Gramineae, Cyperaceae, and *Ephedra*, which together comprise more than 85% of the pollen sum in most pollen samples in the two ice cores.

Although the dominant pollen types are the same, they differ in relative abundance. For example, during 9500 to 6500 BP, a high percentage of Gramineae was found at Guliya, while the percentage of Gramineae was low at Dunde. It is interesting that after 6500 BP the percentage of Gramineae pollen decreased, while an increase in Chenopodiaceae pollen occurred after 6500 BP at Dunde, suggesting that a weakening monsoon occurred at the same time in the two regions.

Ephedra pollen percentages are much higher at Dunde (average >20%) than that at Guliya (<20%). Surface pollen studies from western China (Weng et al., 1993; Van Campo et al., 1996) show that the pollen percentages of *Ephedra* may range from 3% to 80%. In our surface samples from the northeastern Qinghai-Tibetan Plateau, the percentages of *Ephedra* pollen are higher (about 10%) in the desert area than in the steppe and meadow (<5%) region. *Ephedra* commonly occurs in desert and desert-steppe areas (Zheng et al., 1979). The higher percentages of *Ephedra* at Dunde may be due to its proximity to a major desert region (i.e. the Qaidam Basin) to the south and west.

Cedrus pollen was found in the Guliya ice core but is absent in the Dunde ice core. *Cedrus* does not grow naturally in northern Qinghai-Tibetan Plateau. The natural range of *Cedrus* is between 1800 and 2500 m in the western Himalayas (Ministry of Forest, 1981). *Cedrus* can withstand cold winter temperatures (-25°C) but require a humid climate (Sun et al., 1986). Jarvis (1993) found that *Cedrus* pollen was at its highest values from 10,000 to 8500 BP (12,900 to 9000 cal BP) in Sichuan province, Southwest China. Its pollen may have been transported by wind from southwestern Tibet and the Himalayas to Sichuan over a distance of hundreds of kilometers. The occurrence of *Cedrus* pollen during 9500-7500 BP in the Guliya ice core was probably due to an intensified summer monsoon that influenced the northern Qinghai-Tibetan Plateau.

In addition, total pollen concentrations show differences between the two ice cores (Figure 8.5). The average total pollen concentration (1042 grains/liter) in the Guliya ice core for the last 12,000 years is higher than that (412 grains/liter) in the Dunde ice core for the last 11,000 years. The higher pollen concentrations in the Guliya ice

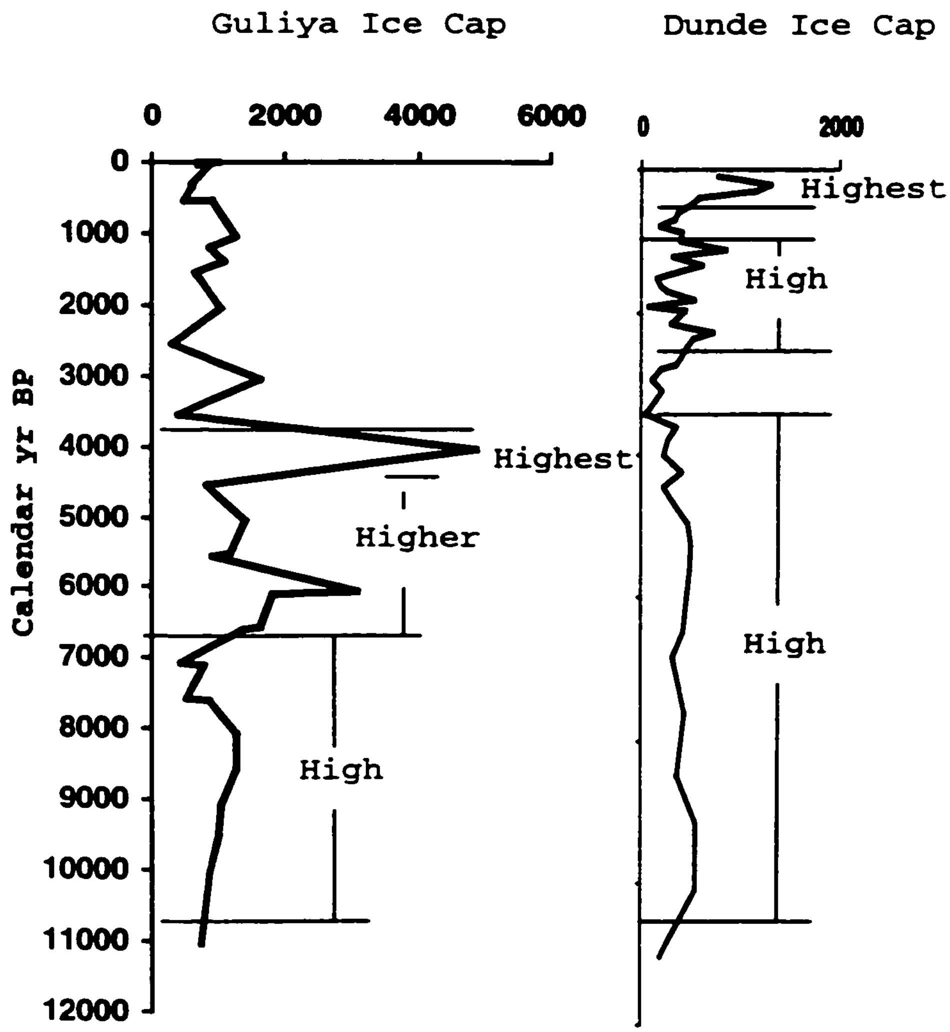


Figure 8.5 Comparison of pollen concentration value (grains per liter) in the Guliya and Dunde ice cores. The highest pollen concentrations occurred at different times at the two ice cores.

core are probably due to the differences in pollen sources, moisture sources, and wind directions between the two ice cores. *Cedrus* and *Abies* pollen were only found in the Guliya ice core, on the other hand, higher percentages of *Ephedra* pollen occurred in the Dunde ice core. Contemporary climatic observations also show that moisture sources and wind directions are different in the two ice cores (Luo and Yanai, 1983; Tang et al., 1979; Thompson et al., 1989).

However, total pollen concentrations show different patterns in the two ice cores (Figure 8.5). At Guliya, pollen concentrations were highest between 6500 and 3900 BP. At Dunde, on the other hand, the pollen concentrations remained constant from 10500 to 4800 BP, then decreased to very low values between 4800 and 2700 BP. The depressed pollen concentration values in the Guliya ice core during the early Holocene (9500-6500 BP) were probably because Gramineae produced less pollen than *Artemisia*. In addition, the maximum pollen concentration occurred at 4500 BP in the Guliya ice core, while it occurred at 350 BP at Dunde. Bourgeois (1986) found that local and exotic pollen concentrations are higher in late-Holocene samples younger than 3000 BP in the Agassiz ice cores. Andreev et al. (1997) found that maximum pollen concentrations occurred at 700 to 800 BP in the Vavilov Ice Cap. Thus, in different ice cores from different parts of the world, pollen concentrations are determined by different sets of environmental factors unique to each site that may affect regional pollen input (such as vegetation density and pollen productivity) and pollen transport and deposition (such as wind directions and strength) at different time in the past.

However, the pollen records show a comparable paleoclimatic history from the two ice cores. The *Artemisia*/Chenopodiaceae (A/C) ratio is a bioclimatic indicator of available atmospheric and/or soil moisture because *Artemisia* pollen comes from the steppe, whereas Chenopodiaceae pollen primarily represents desert (El-Molismany, 1990; Gasse and Van Campo, 1994; Van Campo et al., 1996; K-b Liu et al., 1998). Figure 8.6 shows the similar trends of A/C changes in the two ice cores, even though the A/C ratio probably significantly under-represents the moisture condition between 9500 and 6500 BP at Guliya due to the replacement of *Artemisia* by Gramineae pollen, another moisture indicator. The high A/C ratios suggest maximum monsoon precipitation and high effective moisture. During the late Holocene (after 4700 BP), the decreasing trend of the A/C ratios suggests increasing aridity.

Therefore, the regional climatic changes inferred from the two independent ice-core records correlate remarkably well. Pollen records from the two ice cores suggest that the early to middle Holocene was wetter than at present. This is attributed to the penetration of the summer monsoon to both

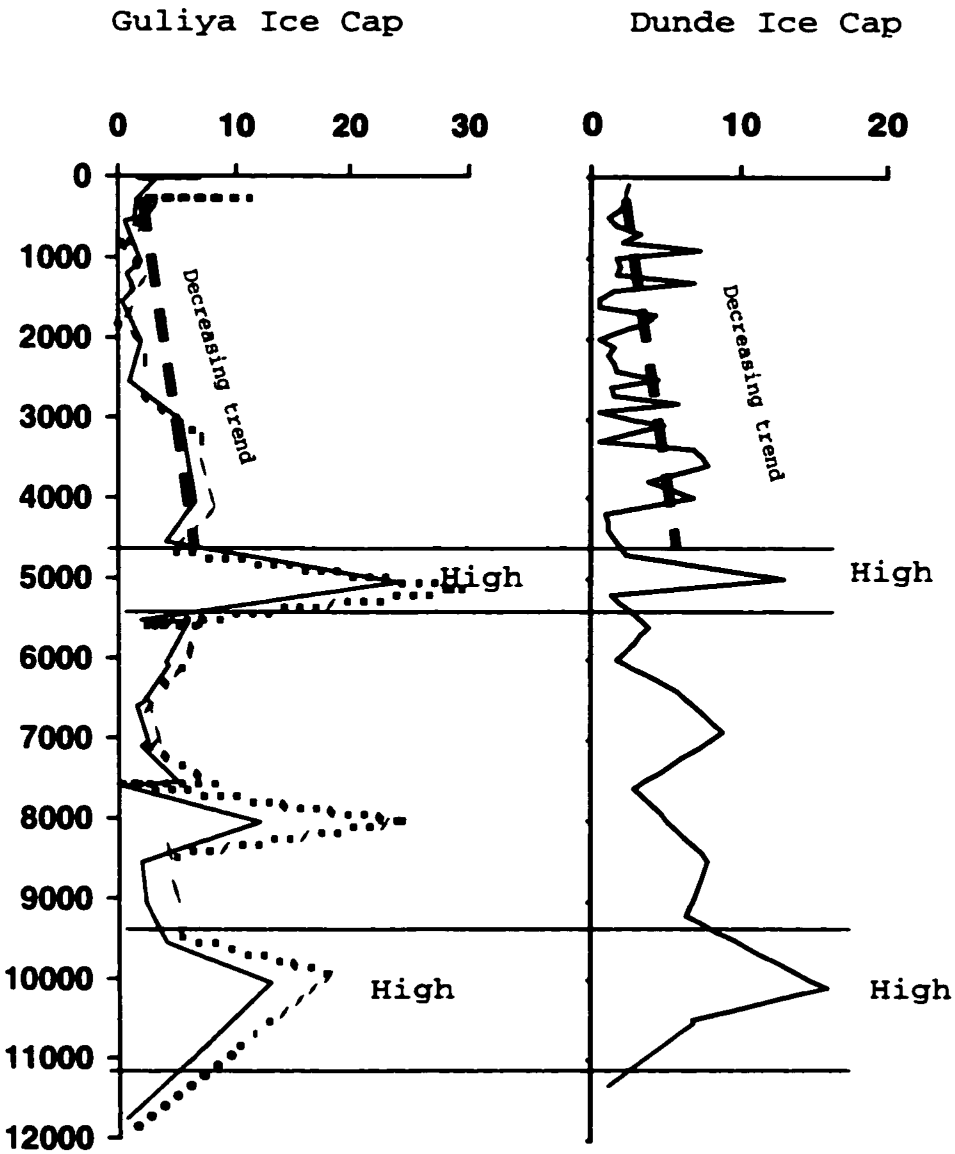


Figure 8.6 Comparison of *Artemisia*/Chenopodiaceae ratio in (a) the Guliya ice core and (b) the Dunde ice core. The dashed line reflects a decreasing trend of the A/C ratio after 4,700 BP. The dotted line represents the ratio of (*Artemisia*+Gramineae)/Chenopodiaceae.

sites. A trend towards aridity started around 6500 BP, characterized by an increase of Chenopodiaceae pollen in the Dunde ice core and a decrease of Gramineae pollen in the Guliya ice core. Several late-Holocene climatic changes, such as the Little Ice Age and Medieval Warm Period (K-b Liu et al., 1998; Chapter VII), occurred in the Dunde ice core, but were not in the Guliya ice core. This is due to the low pollen sampling resolution in the Guliya ice core.

8.7 Comparison with the Sumxi Co Pollen Record

Sumxi Co (34°30'N, 82°23'E, 5058 m, 24.5 km^2) (Number 2 in Figure 8.10) is the closest paleoclimatic study site to the Guliya Ice Cap (Number 9 in Figure 8.10) in the Qinghai-Tibetan Plateau. Sumxi Co is a freshwater lake mainly supplied by the meltwaters of the Mawang Kangri glaciers; the low salinity (0.5 g/l) is attributed to infiltration through the lake floor (Van Campo and Gasse, 1993). Hydrological and climatic changes in Sumxi Co were deduced from the study of a 13000-year core. The detailed pollen and diatom records are discussed in Van Campo and Gasse (1993).

The pollen record from the Guliya ice core differs from that of Sumxi Co in two ways (Figure 8.7). First, the pollen concentration patterns are different. Although it is not possible to directly compare the value of pollen concentrations in an ice core with those in a lake sediment

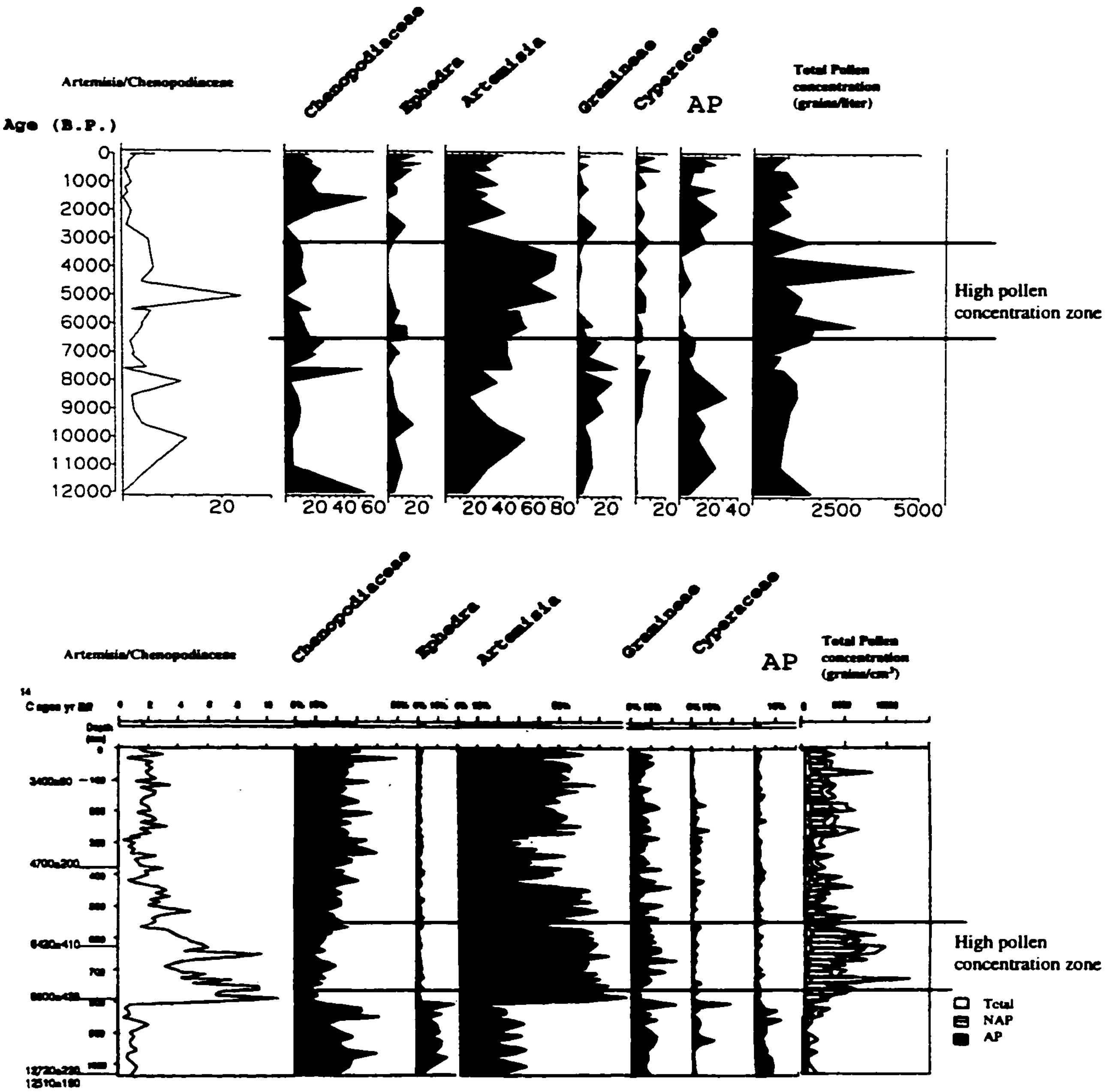


Figure 8.7 Percentage diagram of selected pollen types. (a) the Guliya ice core; (b) Summxi Co (Modified from Van Campo and Gasse, 1993). The highest pollen concentrations are correlated with *Artemisia*-dominated steppe.

core because the processes of pollen deposition and the measurement units are different, the trends of pollen concentration changes can be compared. The higher pollen concentrations occurred from 6500 to 3500 BP in the Guliya ice core, while the pollen record from Sumxi Co shows maximum pollen concentrations from ca 9900 to 7000 BP (cal 11,300-7800 BP) (Figure 8.7). The zone of maximum pollen concentrations in the Guliya ice core lagged behind that in the Sumxi Co core. However, the maximum pollen concentrations all correspond to *Artemisia*-dominated steppe at the both sites, supporting the notion that *Artemisia* produces more pollen than other plants.

Second, the A/C ratios of the two sites also show time lag, while the A/C ratios at Sumxi Co had dropped to low values by 600 BP, the value at Guliya remained high until 3000 BP (Figure 8.8). However, the changes of A/C ratios from Sumxi Co seem to parallel those of Gramineae pollen in the Guliya ice core. Therefore, for the Guliya ice core, the A/C ratio is less useful as a moisture index mainly because four pollen types (Chenopodiaceae, *Artemisia*, Gramineae, and tree) (Figure 8.7) dominate the pollen spectra. Van Campo et al. (1996) also found that the A/C ratio was less useful in the pollen record from Bangong Co (Number 3 in Figure 8.10), a lake near Sumxi Co, because of the co-dominace of several

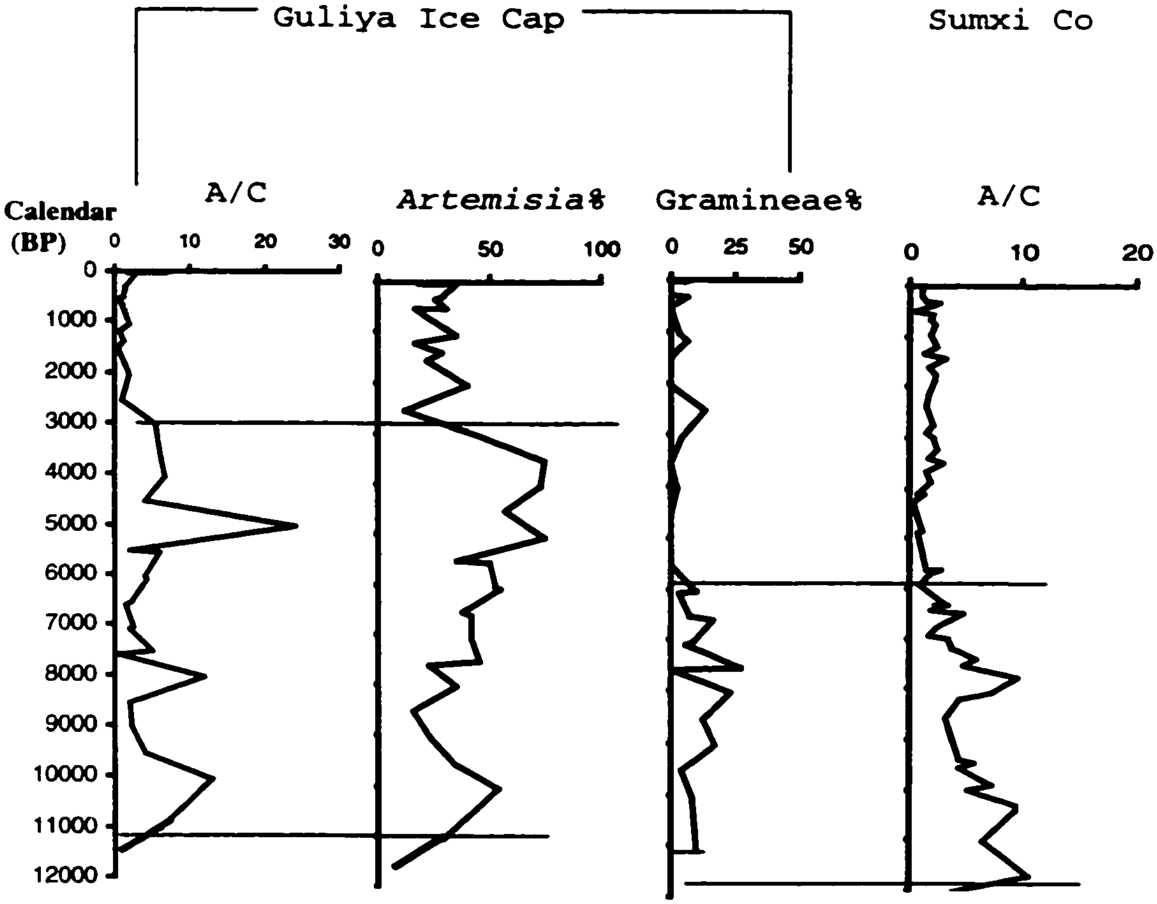


Figure 8.8 Comparison of pollen records between the Guliya ice cap and Sumxi Co. The left three curves are from the Guliya ice cap. The right curve is from Sumxi Co.

pollen types (Chenopodiaceae, *Artemisia*, Gramineae, and Cyperaceae) in the pollen diagram (Figure 2.20).

The difference between an ice-core pollen record and a nearby lake-sediment pollen record is also reported by Andreev et al. (1997), although they did not give an explanation. I propose that the reason for the discrepancy in the pollen diagrams between the Guliya ice core and Sumxi Co is that the two sites have a 1652 m difference in elevation. This results in different pollen source because they are located in different altitudinal vegetation zones. Halpin (1994) found that vegetation zones, situated along an altitudinal gradient, responded differently to climatic change. An increase of 3.6°C in temperature and 10% in precipitation would result in the loss of two representative vegetation zones and the expansion of low and mid altitude vegetation zone (Figure 8.9). Therefore, the discrepancy between the Guliya and Sumxi Co pollen records probably reflects the differential response of different altitudinal vegetation zones to climatic changes.

8.8 Climatic Changes in the Qinghai-Tibetan Plateau

8.8.1 Time-transgressive Nature of the Summer Monsoon

Recent interest in the fluctuations in the strength of the summer monsoon in Africa, India, and China has given rise to various attempts to explain the timing and strength

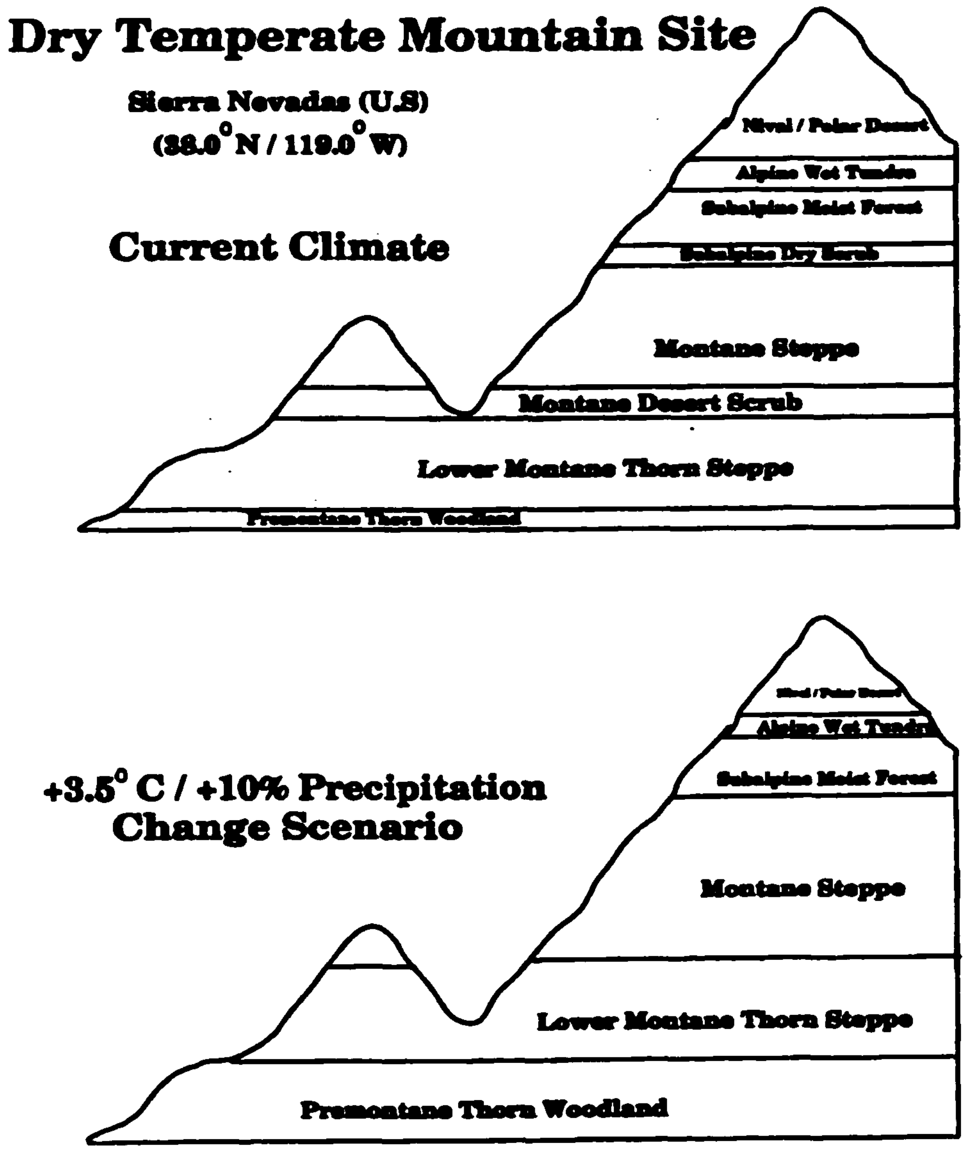
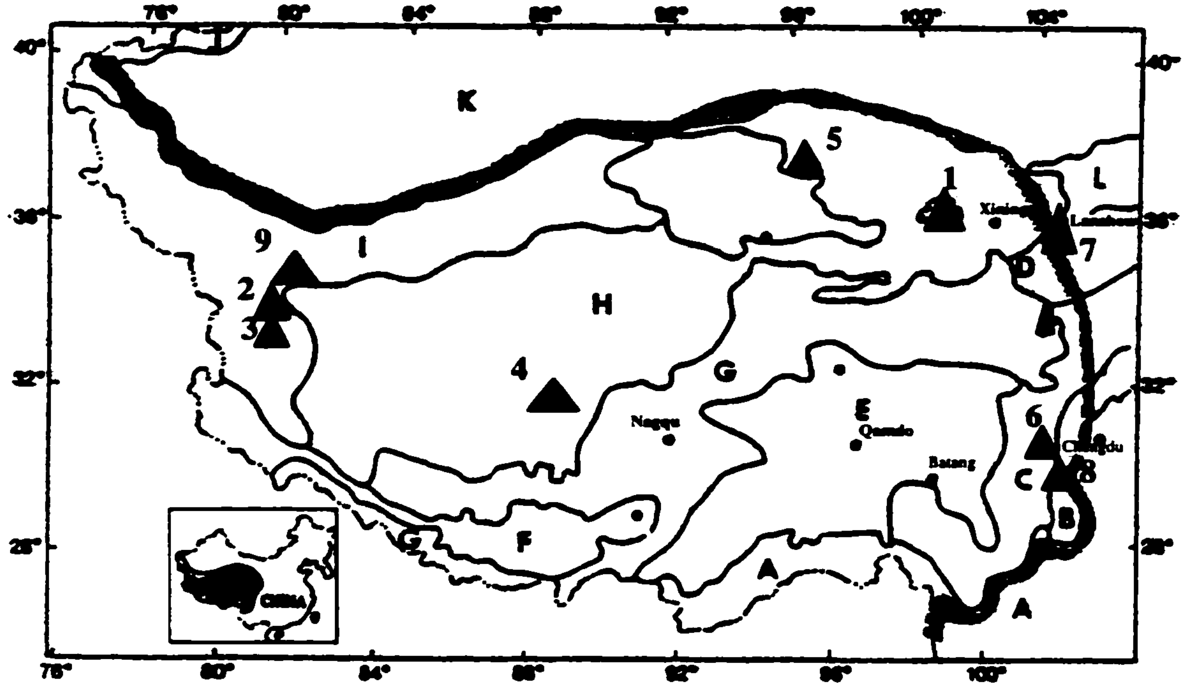


Figure 8.9 Current and changed ecoclimatic zonation for a hypothetical montane temperate climate located in the Sequoia-Kings Canyon Biosphere Reserve, Central California, USA. (After Halpin, 1994).

of the summer monsoon system during the Holocene (Gasse and Van Campo, 1994; Overpeck et al., 1996). These syntheses have depended on two sites in Asia: Western Tibet (Sumxi Co, Bangong Co) and India (Didwana and Linkaransar). However, their works do not include eastern Tibet and eastern China, important areas of modern monsoonal precipitation.

Using GCM, An et al. (1993a) proposed that the monsoon reached maximum strength at different times in different zones within China. Jarvis (1993) reviewed the Chinese palynological literature and pointed out that the evidence was insufficient to support the hypothesis that the onset of the summer monsoon was time-transgressive, because of the relatively few studies from well-dated sites (Liu, 1988).

Intra-plateau patterns of change in the monsoon regime have been discussed (Sun and Chen, 1991; Shen and Tang, 1996; K-b Liu et al., 1998). But earlier works do not consider the role of the elevational gradient in affecting climatic changes. In this study, pollen records from nine sites (Figure 8.10) will be reviewed to test the hypothesis that the front of the summer monsoon maximum moved gradually northward and westward during the early Holocene and retreated from west to east and north to south in the Qinghai-Tibetan Plateau.



A Middle subtropical evergreen broadleaved forest
B Northern subtropical evergreen broadleaved forest
C Warm temperate deciduous broadleaved forest
D Warm temperate dry steppe
E Subalpine forest-meadow
F Subalipne meadow
G Alpine meadow
H Alpine Steppe
I High-cold desert
J Arid mountain vegetation
K Arid desert
L Temperate desert-steppe

Figure 8.10 Location of study sites in relation to vegetation region. (After Jiao, 1984; Liu, 1997). 1 Qinghai Lake; 2 Sumxi Co; 3 Banggong Co; 4 Selincuo; 5 the Dunde ice cap; 6 Lake Shayema; 7 Baoxie; 8 Dahaizi Lake; 9 the Guliya ice cap.

In the central Qinghai-Tibetan Plateau, Sun et al. (1993) reported a pollen record of postglacial vegetation changes at Selincuo (88°31-89°21'E, 31°34-31°57'N, 4530 m a.s.l) (Figure 8.11). An abrupt vegetation change took place at 9600 BP (corrected ^{14}C ages), when alpine sparse vegetation was replaced by alpine grassland. This change lasted until ca. 6000 BP (corrected ^{14}C age). Decreased percentages of *Artemisia* pollen and increased *Pinus* and spores suggest warm and wet condition in this period, implying maximum summer monsoon. During the last 6000 years, the vegetation has remained as alpine grassland, but there has been an increase of arboreal pollen (long distance transport), indicating a decrease of vegetation cover. During the last 1000 years, both the concentration and percentages of arboreal pollen have decreased, suggesting a reduction of forest due to human disturbance.

In southeastern Qinghai-Tibetan Plateau, Jarvis (1993) provided a pollen record from Lake Shayema (~28°40'N, ~102°05'E, 2400 m a.s.l), Sichuan Province (Figure 8.12). She found that cold tolerant species, such as *Abies*, *Betula*, and deciduous oaks dominated the vegetation from 11,000 to 9100 BP. From 9100 to 7800 BP, the abundance of deciduous oaks decreased, and evergreen oaks increased, suggesting a warm and wet climate. From 7800 to 4000 BP, sclerophyllous species increased at the expense of mesic deciduous species,

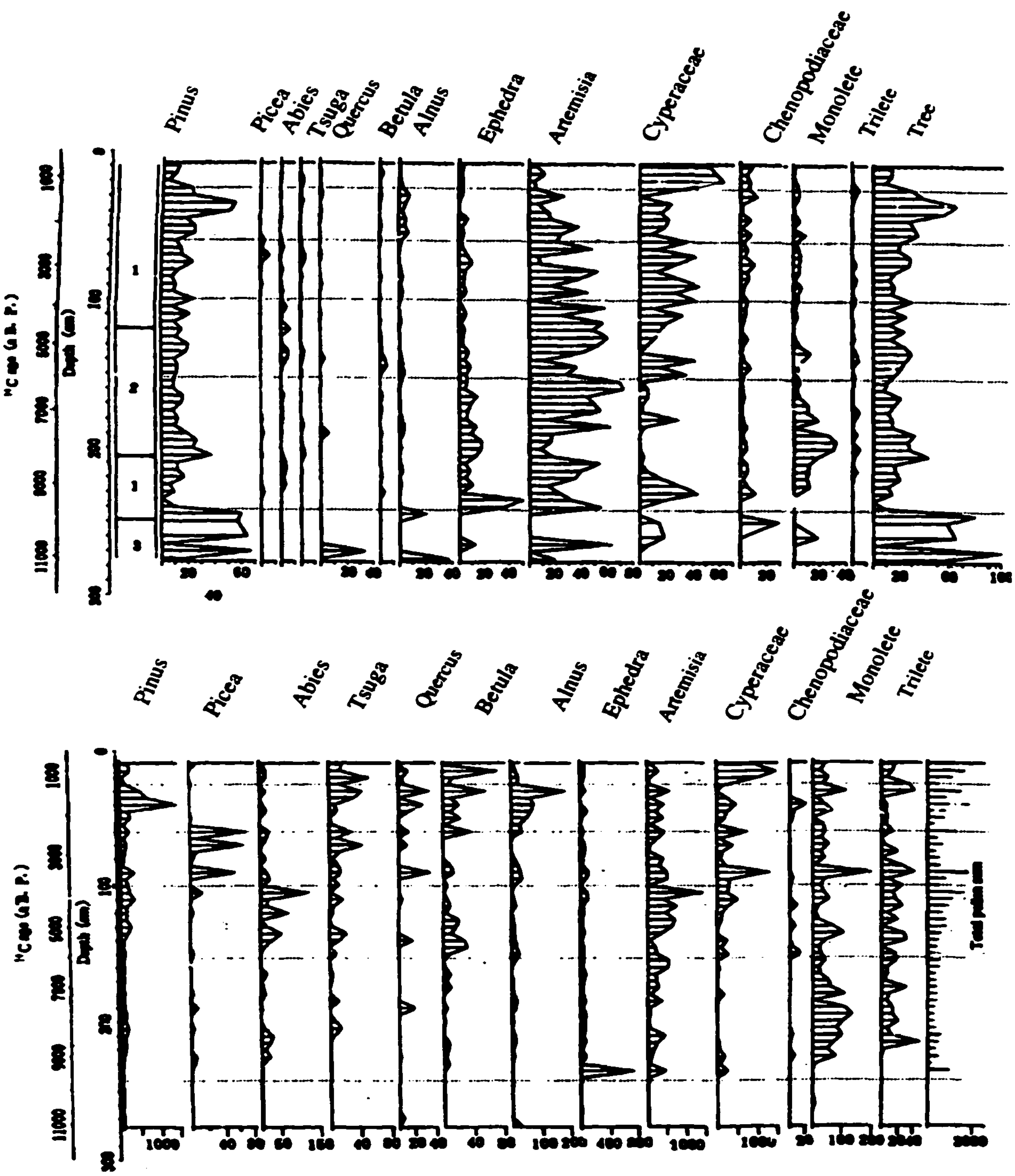


Figure 8.11 Summary pollen diagram for Lake Selincuo. (a) pollen percentages; (b) pollen concentrations. (Modified from Sun et al., 1993).

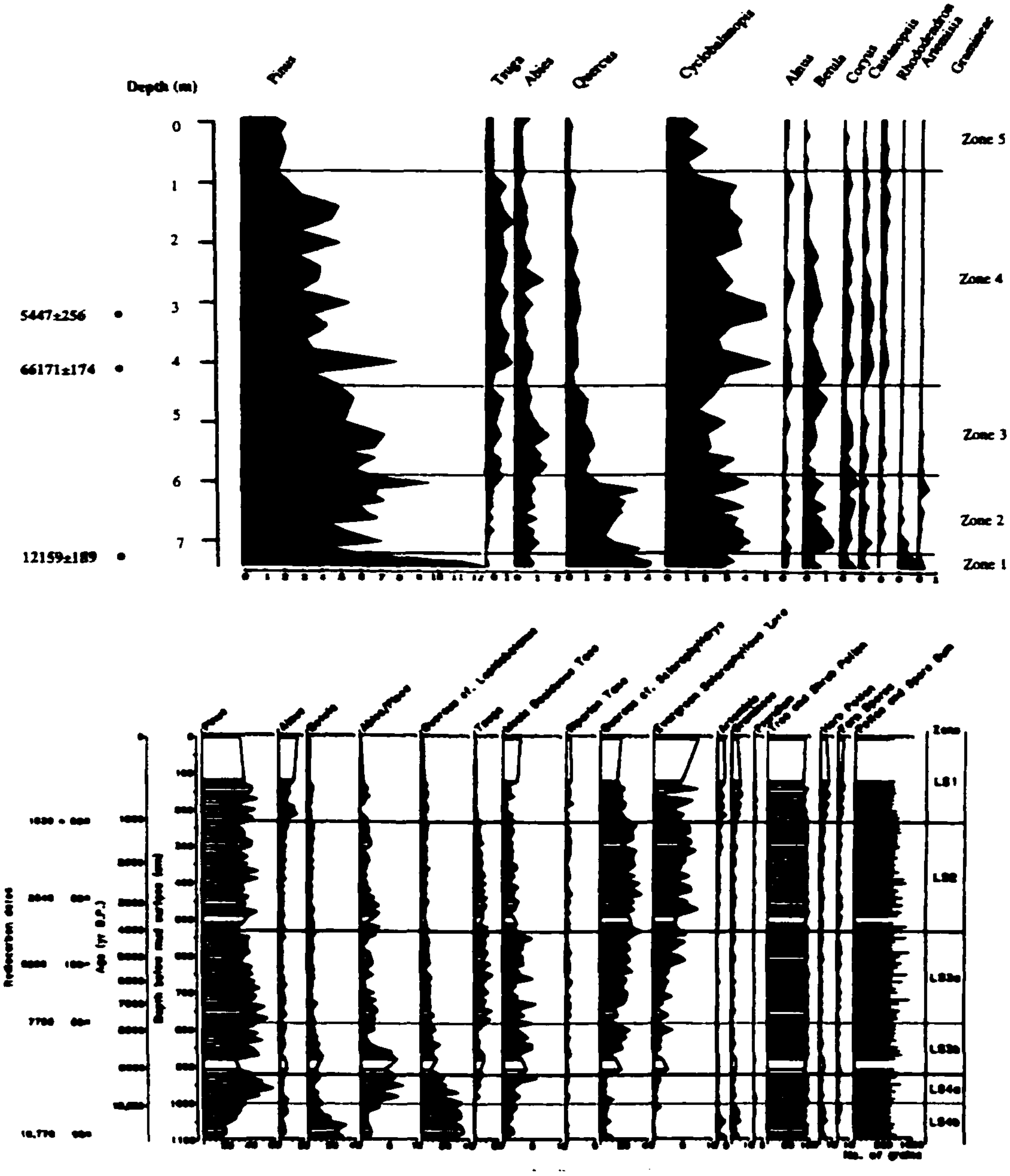


Figure 8.12 (a) pollen concentration diagram from Lake Dahazhi (After Li and Liu, 1988); (b) pollen percentage diagram form Lake Shayema (After Jarvis, 1993).

an indication that precipitation was becoming more seasonal. Increased disturbance started ca 1000 BP. Li and Liu (1988) analyzed pollen data from Dahaizi Lake (27°30'N, 102°24'E, 3660 m a.s.l) on Luoji Mountain, approximately 100 km south of Shayema Lake. Li's data recorded the replacement of deciduous oak pollen by that of evergreen oak and the gradual replacement of conifers and deciduous broadleaved taxa by evergreen broadleaved sclerophyllous taxa beginning ca. 7600 BP and lasting until 2000 BP. They did not, however, have good dating control: the boundaries of vegetational change were estimated.

In the northeastern Qinghai-Tibetan Plateau, Du et al. (1989) provided a pollen record from Qinghai Lake (36°33'-37°15'E, 99°36'-100°47'N, 3196 m a.s.l). Between 10000 and 8000 BP, the vegetation changed from temperate steppe to temperate forest steppe, suggesting the onset of moister conditions (Figure 2.19). Between 8000 and 3500 BP, an increase of tree pollen probably indicates an increase of moisture. However, a decrease of *Betula* and an increase of Chenopodiacae pollen after ca 6000 BP may indicate a decrease of moisture (Kong et al., 1990). This can be supported by evidence of lake level change. The level of Qinghai Lake was highest at 7000 BP and fell gradually by about 10 meter to its present level by ca 3000 BP (Kelts et al., 1989), probably suggesting a weakened summer monsoon.

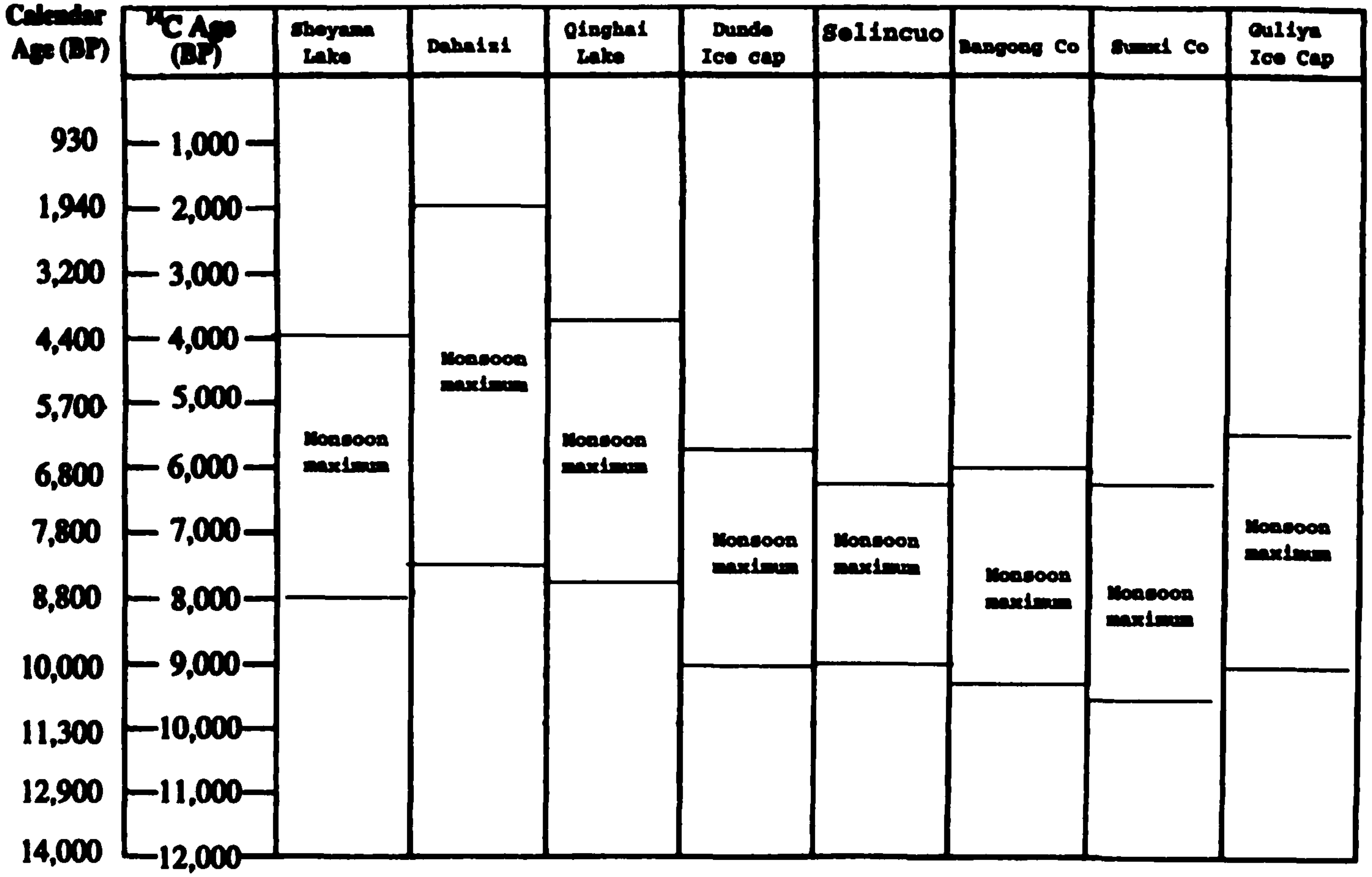
From 3000 to 1500 BP, arboreal pollen was still significant, but *Artemisia* recovered to pre-Holocene values. A pollen record from the Dunde Ice Cap (Figure 2.18), west of the Qinghai Lake, shows a similar trend (K-b Liu et al., 1998). High pollen concentration occurred between 10,500 and 4800 BP (calendar years), suggesting wet condition. But the increase of Chenopodiaceae and *Ephedra* pollen beginning 6500 BP probably suggested a weakened summer monsoon until 4800 BP, following the driest period between 3500 and 2900 BP. A paleosol study from Baxie (35°33'N, 103°35'E; 2000 m a.s.l) in the Loess Plateau, east of Qinghai Lake, reveals the development of a humid-climate paleosol at ca 11,000 BP, followed by a reversal to drier conditions at ca 9800-9300 BP (An et al., 1993b). A longer episode of paleosol development occurred between 9300 BP and 5500 BP, indicating a period of strongest summer monsoon. After 5500 BP (6300 cal BP) loess deposition, probably correlated with an increase of Chenopodiaceae pollen at Dunde, indicates a dry condition.

In western and northwestern Qinghai-Tibetan Plateau, pollen studies from the Sumxi Co and Bangong Co provided a continuous Holocene pollen record (Figure 2.20, 2.21). At about 9600 BP, the dominance of desert elements in the pollen record abruptly shifted to steppe vegetation, suggesting the onset of moister conditions (Van Campo et al.

1996). This wet condition lasted until ca 6000 BP (^{14}C age BP). Short-term oscillations are superimposed on this general trend of climatic change, with a major reversal event at about 8000 BP at Sumxi Co and at 7700 BP at Bangong Co. After 6000 BP, a trend towards aridity is observed with dry events centered around 5500, 3900-3200, and 700 BP. A new pollen record from the Guliya Ice Cap (this study) provides a comparable history of climatic changes. Before 9500 BP, vegetation is dominated by *Artemisia* and Chenopodiaceae. An increase of Gramineae pollen beginning at 9500 BP, accompanied with high charcoal concentrations, suggests wetter conditions than before. The decrease of Gramineae at 6500 BP suggests that the summer monsoon began to retreat. Between 6500 and 2900 BP, an increase of *Artemisia* indicates drier conditions.

In summary, all the sites show that climatic conditions wetter than those of today prevailed during the early to middle Holocene period. The evidence (Table 8.2; Figure 8.13) is insufficient to support the hypothesis that the front of the summer monsoon maximum moved gradually to north and west during early Holocene and retreated eastward and southward in the Qinghai-Tibetan Plateau. On the contrary, the evidence seems to show that the beginning of the summer monsoon maximum occurred earlier in the western Qinghai-Tibetan Plateau than in the eastern Qinghai-Tibetan Plateau.

Table 8.2 Chronology of inferred summer monsoon maximum for selected sites.



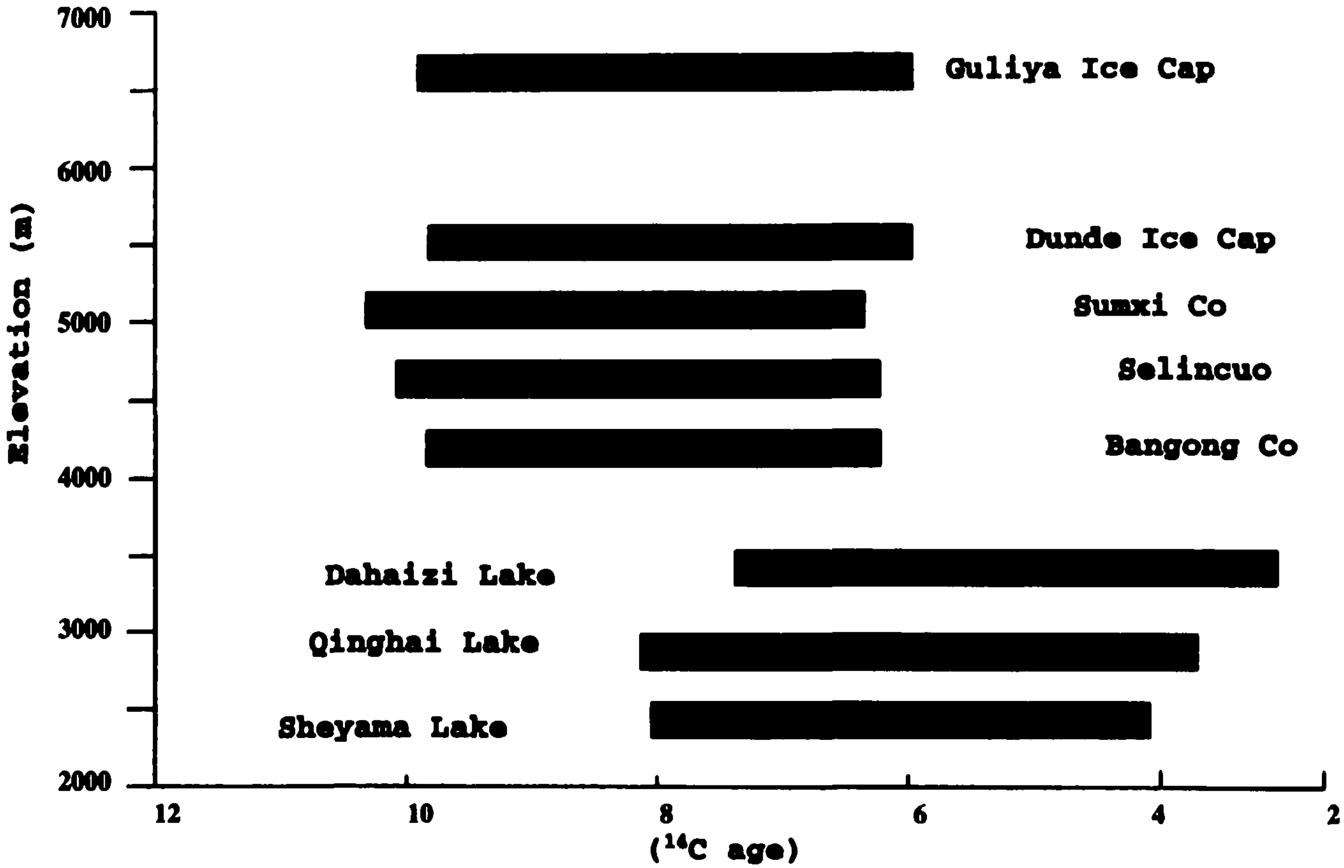


Figure 8.13 The beginning and ending of summer monsoon maximum at different sites. The y-axis represents elevation (m) and the x-axis is the age.

It probably indicates that the two areas were influenced by different monsoon systems. During the summer monsoon maximum, the eastern most part of the Qinghai-Tibetan Plateau was probably influenced at least partly by the southeast monsoon, while the western and central part of the Qinghai-Tibetan Plateau was dominated by the southwest monsoon system. More pollen records from the central and southern parts of the Tibetan Plateau are needed to test this hypothesis.

8.8.2 Climatic Fluctuations during the Early to Middle Holocene

Climatic fluctuations are clearly indicated for the Qinghai-Tibetan plateau during the early to middle Holocene, although the climate during this period was generally wet and warm. The early to middle Holocene wet episode was interrupted by several periods of reduced rainfall (Gasse and Van Campo, 1994). Gasse and Van Campo (1994) reviewed pollen and lake records from West Asia, East Africa (Ethiopia), and West Africa (Western Sahara, Sahel and subequatorial Africa). They proposed that three major dry episodes occurred during the intervals ca 11,000-9500 BP, ca 8000-7000 BP, and ca 4000-3000 BP.

A pollen record from the Guliya ice core (this study) shows that an increase in Chenopodiaceae pollen percentages and a decrease in total pollen concentrations occurred at 7500 BP (8000-7000 BP ^{14}C age), probably correlated with the second major dry period identified by Gasse and Van Campo (1994). The lowest pollen concentration occurred at 3500 BP, probably correlated with the third major dry period identified by Gasse and Van Campo (1994).

Loess deposited in the Kunlun Mountain (close to the Guliya Ice Cap) has provided evidence of three humid periods in this area interrupted by dry periods (Gao and Zhang, 1991). Three thick soil layers are ^{14}C dated at 7080, 5338 and 2486 BP, respectively. Li et al. (1985) found that peatlands had started to develop after 10000 BP in the broad valleys of the Qinghai-Tibetan Plateau above 4500 m. These peatlands had basal radiocarbon dates of 9970±135, 7670±250, and 3575±80 BP. Peatland development declined in the late Holocene (after about 3000 BP) on the Qinghai-Tibetan Plateau and ceased altogether at some sites. In the Loess Plateau, Jiuzhoutai mountain, just north of Lanzhou City, has three palaeosols with ages of 10,000 to 8500, 7500 to 5000, and 2700 to 2400 BP (Zhou et al. 1991). All these suggest that decade- to century-scale variations in monsoon strength occurred in western China.

8.9 Conclusions

The pollen records show that the early to middle Holocene (9500-6500 BP) was warmer and wetter than today in the Qinghai-Tibetan Plateau. This warm and wet period was followed by a dry period from 6500 to 2900 BP. After a transitional period of increased humidity, the modern climatic pattern of pronounced aridity has been established since 1900 BP. The higher charcoal concentrations from 9500 to 6500 BP occurred at a time of maximum humidity in the Qinghai-Tibetan plateau, when a grass-dominated steppe prevailed near the Guliya Ice Cap. This vegetation type, with a more continuous vegetation cover, was more conducive to the generation and spread of wild fires under arid to semi-arid conditions. The charcoal record from the Guliya Ice Cap is, to my knowledge, the first charcoal study in an ice core.

A comparison of pollen records between Guliya and Dunde indicates broadly similar trends of climatic changes. A comparison between the Guliya and Sumxi Co pollen records suggests that the pollen response to climatic changes varies because vegetation zones expanded and contracted at different elevations.

A review of pollen studies from the Qinghai-Tibetan Plateau suggests that the summer monsoon influence had begun by 10,000 BP (^{14}C age). Pollen records from Lake Manas and Lake Aibi in northern Xinjiang (Sun et al., 1994; Li, 1996; Rhodes et al., 1996) suggests that the Indian and Asian monsoons penetrated inland during the early to middle Holocene. Due to inadequate dating control, evidence from the Qinghai-Tibetan Plateau is insufficient to support the hypothesis of time-transgressive monsoon maximum. The early to middle Holocene wet and warm period was interrupted by several dry periods, suggesting century-scale variations in past monsoon strength.

CHAPTER IX

SUMMARIES AND CONCLUSIONS

This research has provided detailed pollen records from two ice caps in the Qinghai-Tibetan Plateau: Guliya and Dunde. In order to establish the relationship between pollen and climate, the modern pollen rain in the northeastern Qinghai-Tibetan Plateau and a 30-year pollen record have been studied. A detailed history of climatic changes at decadal and centennial time scales has been derived from a 2000-year pollen record from the Dunde ice core. A detail monsoon history at millennial time-scale has been recovered from a 12,000-year pollen record from the Guliya Ice Cap. In this final chapter, a summary of the main findings is provided. Future research directions are also discussed.

9.1 Summary of Main Research Findings

9.1.1 The Modern Pollen Rain

In order to reconstruct changes in past vegetation, it is important to determine how modern vegetation types are represented by pollen rain. Eighty-seven surface samples were studied from southeast to northwest of the Dunde Ice Cap. It can be concluded that major vegetation types of the northern Qinghai-Tibetan Plateau have distinctive palynological signatures. Most of the pollen come from local vegetation. The A/C ratio decreased with decreasing precipitation, suggesting that the A/C ratio can be used as a moisture indicator.

9.1.2 The Relationship between Pollen and Climate

This is the first study of the relationship between ice-core pollen and modern climatic observations. The total pollen concentration, percentages of meadow pollen, and percentages of steppe pollen are positively correlated with annual precipitation, summer precipitation, and winter temperature, while they are negatively correlated with summer temperature. This indicates that all these variables can be used as indicators of moisture. The percentages of tree and desert pollen are positively correlated with annual temperature and summer temperature, while they are negatively correlated with annual precipitation and summer precipitation.

Therefore, the total pollen concentration can be considered as a combination of temperature and precipitation because climatic parameters have interactive effects on pollen. The increase of total pollen concentration may imply an increase in moisture or decrease in temperature. The increase of meadow pollen may be an indicator of moisture increase, while an increase of desert pollen may be considered as an indicator of moisture decrease. Higher percentage of tree pollen implies exotic pollen from long-distance transport or from increased temperature.

9.1.3 Climatic Changes at the Decadal and Centennial Time Scales

A 2000-year pollen record from the Dunde ice core provides an indication of decadal and century time scale climatic changes. The pollen data from the Dunde ice core during A.D. 0-280 are characterized by relatively high pollen concentration and high percentages of *Artemisia*, indicating a wet and cold period. Low pollen concentration and high indeterminable pollen indicate a dry and cold period between A.D. 280-640. Following this period, increased pollen concentrations and Cyperaceae pollen percentages indicate a warm and wet climate condition between A.D. 640 and 1200. From A.D. 1200 to 1370, the lowest pollen concentration and highest indeterminable pollen percentages suggest a warm and dry condition, probably correlated with late European Medieval Warm Period. The highest total pollen concentrations occurred during 1660-1910 A.D., implying the prevalence of cool, humid conditions during the Little Ice Age. A dry trend has occurred during the last 80 years.

Moisture variation derived from the pollen records at Dunde during the last 2000 years is similar to that from the Guliya Ice Cap (west of Dunde) and from the semi-arid area of western China (east of Dunde). At least two cold periods occurred on the Qinghai-Tibetan Plateau. One of them is correlated with the Little Ice Age. Therefore, the ice caps on the Qinghai-Tibetan Plateau are sensitive to climatic changes.

9.1.4 The Last 12,000 Years of Climatic Changes

The pollen record from the Guliya ice core shows that the early to middle Holocene is warmer and wetter than today in the Qinghai-Tibetan Plateau. This was followed by a relatively dry period from 6500 to 2900 BP. The modern climatic patterns have been established since 1900 BP in the Guliya Ice Cap. The higher charcoal concentrations from 9500 to 6500 BP suggest frequent wild fires, which was favored by the more continuous vegetation cover in a grass-dominated steppe, a vegetation type developed during a warm and wet period due to an intensified summer monsoon. The charcoal record from Guliya is the first charcoal study in an ice core.

A review of pollen records from nine sites in the Qinghai-Tibetan Plateau indicates that evidence is insufficient to support the hypothesis of time-transgressive summer monsoon maximum. This study also suggests that the early to middle Holocene wet and warm period is interrupted by several dry episodes, implying a century time-scale variation of the summer monsoon.

9.2 Directions for Future Research

The research I have conducted shows the potential for using ice-core pollen data to understand past climatic changes in the Qinhai-Tibetan Plateau. Undoubtedly the quality of interpretation could have been improved with access to more information. The future studies that need to be conducted are briefly discussed below.

A quantitative relationship between pollen data and climate is needed in association with other climatic data such as wind direction and air pressure. The testing of hypotheses such as the variability of the monsoon system depends on a large spatial data network. However, the accurate dating of lake-sediments is affected by the presence of old carbonates, a problem common in Tibetan Plateau lakes. Therefore, good dating control, especially at key climatic boundaries, is needed in the future. An ice-core study is a good way to achieve this, because ice cores usually have reliable dating control provided by the annual layers at least for the last several thousand years. Therefore, pollen studies of ice cores, especially from southern and central Qinghai-Tibetan Plateau, are very useful for the reconstruction of climate changes in central Asia, particularly for the study of monsoon climatic variations.

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DOCTORAL EXAMINATION AND DISSERTATION REPORT

Candidate: Zuju Yao

Major Field: Geography

Title of Dissertation:

Ice-core Pollen Studies from the Dunde and Guliya Ice Caps, Qinghai-Tibetan Plateau, China

Approved:

Major Professor and Chairman

Dean of the Graduate School

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